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GEOLOGY OF THE MARION LAKE AREA, QUEBEC-LABRADOR

by

John Allan Donaldson

A dissertation

submitted to the Faculty of Philosophy

of The Johns Hopkins University in conformity with

the requirements for the degree of

Doctor of Philosophy.

Baltimore, Maryland

1960

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ABSTRACT

The Marion Lake area is in the eastern plateau region of the Canadian Shield. Schefferville, center of the Quebec-Labrador iron range, is about 35 miles west of Marion Lake.

The eastern one-third of the area is underlain by amphibolites and quartzo-feldspathic gneisses. The western two-thirds is underlain by metabasalt, metagabbro, serpentinite, and slightly metamorphosed sedimentary rocks of the Labrador Trough, a narrow belt of Proterozoic rocks which extends approximately 450 miles northwest and 125 miles southeast of the Marion Lake area.

The gneisses and amphibolites form an eastern block which truncates northwest-trending folds and faults in the western block of Trough strata. The north-trending contact is a reverse fault, along which the gneissic block moved up relative to the western block. The gneisses and amphibolites are either part of the Archean basement or metamorphosed Trough rocks.

Sedimentary strata of the Knob Lake group, with a possible total thickness of more than 20,000 feet, have been mapped in the southwestern corner of the area. A northwest-trending fault separates these rocks from the volcanic belt in the central part of the area. Greenschists of the Murdock group crop out immediately east of the fault. The overlying Doublet group includes 3500 feet of sedimentary rocks, and more than 10,000 feet of metavolcanic rocks.

Remarkably conformable basic sills underlie much of the area. The volcanic and intrusive rocks were probably derived from the same magma. The sills, which are as much as 5000 feet thick, were intruded prior to, and during the early stages of, folding.

Most sedimentary rocks in the area show evidence of deposition on a stable shelf which was, at most times, covered by shallow water. Graded bedding, probably formed as a result of the annual climatic cycle, is exhibited by many of the fine-grained sedimentary rocks. Cross-bedding attitudes in sandstones beds west of Marion Lake indicate sediment transport from the west.

The Denault formation of the Knob Lake group, which was studied in detail, is composed of approximately 30 percent finely laminated siliceous dolomite, 20 percent massive dolomite, 5 percent intraformational conglomerate, and 45 percent stromatolitic dolomite. Five zones of distinctive stromatolites were correlated in sections more than two miles apart. Algal activity was important in trapping sediment, and may have played a dominant role in primary precipitation of dolomite.

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INTRODUCTION

Regional Setting

This dissertation describes an area of some 360 square miles at the eastern margin of the Labrador Trough, a belt of extensively folded and faulted metasedimentary, metavolcanic, and basic intrusive rocks of Proterozoic age. The belt extends southeast from the west side of Ungava Bay for a distance of more than 500 miles (Fig. 1).

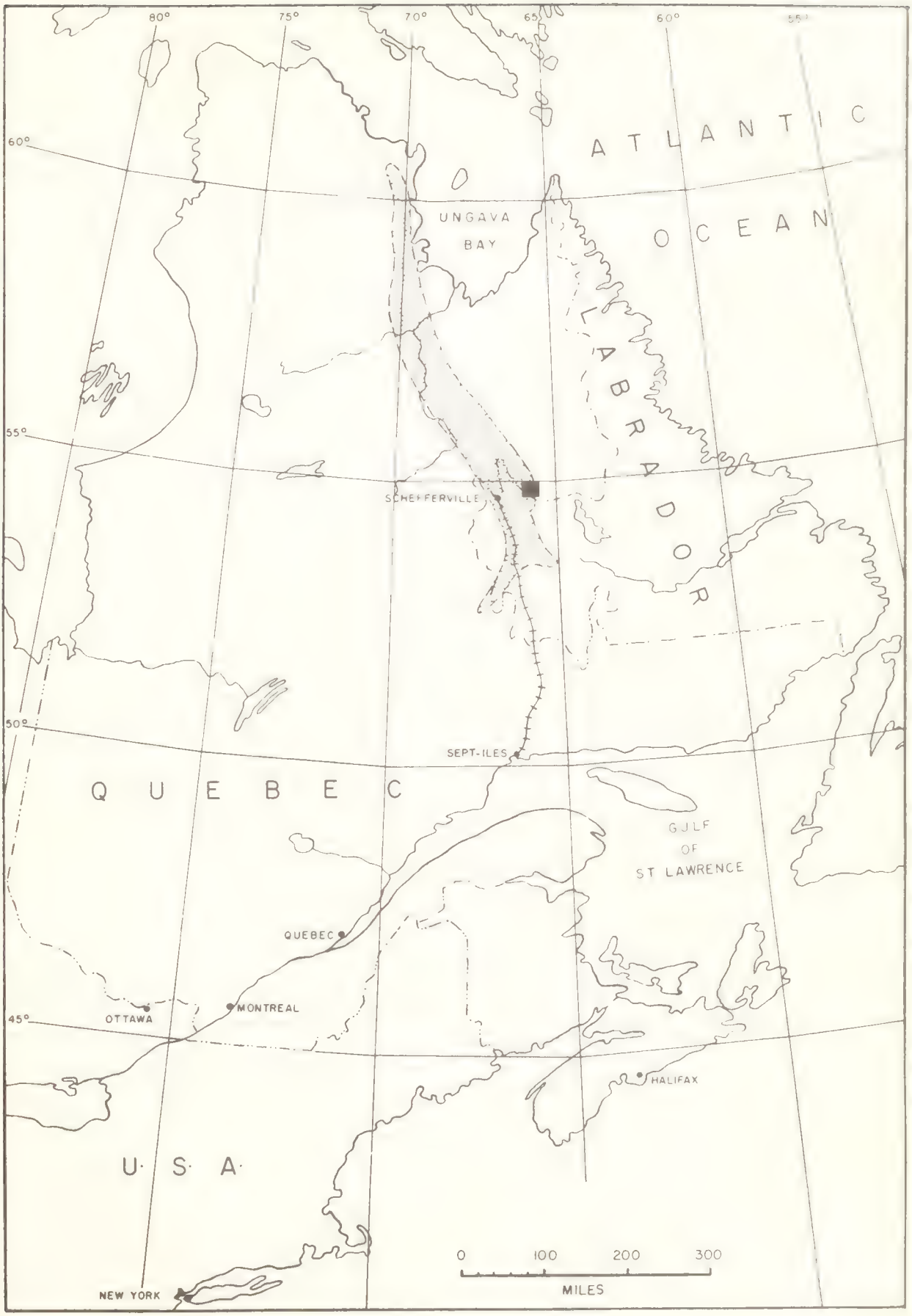
In general, sedimentary rocks underlie the western portion of the Labrador Trough, and volcanic rocks underlie the eastern portion. The western boundary of the Trough is a major unconformity (Harrison, 1952, p. 3); the eastern boundary is a zone of extensive faulting (Fahrig, 1957, p. 116). The degree of metamorphism increases from west to east in the northern part of the Trough (Sauvé, 1957, p. 153), and in the southern part of the Trough, Duffell (1959) has mapped high-grade metamorphic equivalents of the major rock types which occur farther north.

Location and Access

The Marion Lake map-area, bounded by latitudes 55°00' and 54°45' N and longitudes 66°00' and 65°30' W, comprises two adjacent 15 minute quadrangles. The area is about 35 miles east of Schefferville, center of iron mining activity in Quebec and Labrador. Quebecair has scheduled flights to Schefferville, and the Quebec North Shore and Labrador Railway provides passenger service via the 360 mile railroad between Seven Islands and Schefferville. During the field season, sea-planes may be chartered at Squaw Lake, two miles north of Schefferville. There are no good water routes to Marion Lake, but canoes greatly facilitate work within the area.

Fig. 1

Index map, showing location of Labrador Trough (stippled), and the Marion Lake area (black rectangular area).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MILES

Previous Work

A. P. Low (1895) conducted the first geological investigations in the Labrador Trough. On one of his reconnaissance traverses, he reached the headwaters of the Ashuanipi River system, 20 miles west of Marion Lake. Since the discovery of iron ore by Gill at Ruth Lake in 1929, much exploration and development work has been carried out within the Labrador Trough. Effort was at first directed towards development of the iron ore deposits near Schefferville, but sulphide and asbestos prospects along the east side of the Trough have recently attracted attention. The Labrador Mining and Exploration Company Limited has mapped parts of the Marion Lake area, but results of their work have not been published. The Willbob Lake area (Frarey, 1952)*, to the northwest and the Griffis Lake area (Fahrig, 1951)*, to the north have been mapped on a scale of one-half mile to one inch. Frarey mapped the Menihek sheet to the west on a scale of four miles to one inch, but it has not yet been published. The areas to the south and east are unmapped.

Field Work

Mapping was carried out during the periods June 29 to August 26, 1957 and June 27 to September 8, 1958. Royal Canadian Air Force aerial photographs on a scale of approximately one-half mile to one inch were used in the field. Base-maps on a scale of 1:40,000, provided by the Topographical Survey, Department of Mines and Technical Surveys, were used for compilation of field data. One of these maps, enlarged to a scale of 1:31,680, served as a base for the final drafting.

Traverses were spaced at intervals of one-half mile or less,

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*See index map, in pocket at back, for location of other areas.

and, in most cases, were directed across the strike of the rock units. In the heavily drift-covered eastern part of the area, they were adjusted to cross all outcrops visible on the aerial photographs. In the western portion of the area, where exposures are abundant, traverses were planned to allow the study of critical areas. Outcrops were located by pace and compass methods, and turning points were taken only at positions which could be definitely identified on the aerial photographs. Traverse lines were tied in to streams, lake shores, or other topographic features, and the pace corrected to accord with actual distances. Parts of the well-exposed Marion Lake and Denault formations were studied in detail. Five stratigraphic sections were measured in the Denault formation and specimens were collected along each section line.

Laboratory Work

The laboratory work was done at The Johns Hopkins University during the academic years 1957-58 and 1958-59. Thin sections of 130 specimens, representative of the lithologic units in the Marion Lake area, were studied. The universal stage was used (Emmons, 1943) to determine the compositions of feldspars in the gneisses and amphibolites, but most mineral identifications were made with the flat-stage petrographic microscope. The Norelco x-ray diffractometer provided identification of minerals in a few specimens of dolomite, sandstone and slate.

Modal compositions were determined by means of the point-counter (Glagolev, 1933; Chayes, 1949). Sufficient points were counted for each analysed thin section to obtain frequency distributions which would not be appreciably changed by additional data. Average roundness

and sphericity of detrital grains were estimated by comparison with the chart figured by Krumbein and Sloss (1953, p. 81).

A microprojector was used for thin-section size analyses of the sandstones. The apparent long axes of all sand grains in a random field of projection were measured, and the data grouped in size classes. This procedure was repeated until measurements of at least 400 grains were obtained for each thin section. Conversion of the number frequencies to approximate volume percent frequencies by the method described in Appendix A allows comparison with size distributions of modern sands.

The microprojector method was also used for size analyses of the sand fractions of conglomerates, but because a thin section provides too small a sample of pebbles and granules, the number frequency distributions of these larger particles were determined by direct measurement on polished surfaces. Results of the analyses were combined in proportion to relative surface areas of the sand, gravel, and matrix fractions.

Number frequency distributions and calculated volume frequency distributions for sandstones and conglomerates are listed in Appendix B.

For detailed study of dolomite structures and textures, specimens were cut perpendicular to bedding. One cut surface of each specimen was ground to a smooth finish with carborundum powder, and the matching surface was etched in a 10 percent solution of hydrochloric acid. The staining technique described by Hobbs (1957, p. 17) was used to check for the presence of calcite, and the spot test outlined by Mann (1951), was used to detect phosphates.

Eleven samples, each consisting of fresh chips collected across strike of the outcrop sampled, were reduced to -35 mesh and submitted to the Geological Survey of Canada for chemical analysis. The

samples were analyzed by means of a rapid method which gives a total calculation of 99.5 to 100.5 percent.

Physiography

The Marion Lake area is crossed by the height of land marking the unsurveyed provincial boundary between Labrador and Quebec. Gauthier, Moss, and Marion Lakes, on the Labrador side of the watershed, are headwaters of the McKenzie River, which drains southeast to the Atlantic Ocean via Hamilton River. Rivière de Pas drains the Quebec side northward to Ungava Bay via the George River.

Topography is closely related to lithology and structure of the bedrock. Folded and faulted rocks of the Labrador Trough underlie numerous hills and ridges which rise 400 to 600 feet above intervening lakes in the western two-thirds of the area; gneisses and amphibolites underlie comparatively flat and swampy terrain in the eastern third. Lakes occupy nearly one-quarter of the area, and muskeg covers almost half of the remaining area. The larger lakes have elevations between 1500 and 1600 feet, and the highest hill, just west of Lac Tantouin, is 2091 feet above sea level.

Black spruce are the most abundant trees in the area. Some white spruce and balsam grow in the thicker stands of timber, and tamarack is abundant at the margins of lakes and swamps. In protected valleys, the trees are commonly more than 100 feet high, with butt diameters up to 30 inches, but on wind-blown slopes they seldom exceed a height of 50 feet. Willows, alders, and small shrubs cover the sides of hills and form dense thickets along stream channels. The ubiquitous arctic moss is, in many places, the only bedrock covering.

The climate of the area is sub-arctic and the weather is generally unsettled. More than half the days of the 1957 and 1958 field seasons were overcast, windy, and rainy. Summer temperatures seldom drop below freezing or rise above 80 degrees, but fluctuate abruptly between these limits. Some snow remains until early July, and may be expected again in early September. Most lakes are free of ice by the end of June, and seldom freeze over until mid-September. The mean annual temperature is approximately 23° Farenheit and the annual precipitation exceeds 30 inches.

Pleistocene Geology

Glacial striations and grooves are visible on the surface of many rock outcrops in the area, especially those exposed on high hills. At least two different directions of regional ice-movement are indicated by sets of striations (Plate 1), which are present on widely separated outcrops. The strike of the youngest and most prominent set ranges from 49° to 55°. Roches moutonées, drumlin configurations, and plucking at the ends of grooves indicate that the ice-sheet advanced from the southwest. The earlier set of striations, partially obliterated by the later movement, strikes at 115° to 130°, parallel to the trend of ridges in the Labrador Trough. Ice-movement for this set was from the northwest.

Glacial drift in the area consists mostly of boulder till, but there are extensive sandy outwash deposits around Marion Lake and east of Lac Villedonné. In the eastern part of the area, where the drift has a high clay content, there are many small drumlins and larger drumlin-like ridges. Only the small drumlins can be recognized easily in the field; the larger forms, particularly abundant east of Lac

PLATE 1

Glacially polished bedrock

Outcrop showing striations and plucking due to Pleistocene glaciation. Ice movement was from the left, and parallel to the hammer handle.

Gouffier, are features clearly seen in aerial photographs. They are as much as 50 feet high and 3000 feet long, and their long axes are parallel to striations and grooves formed by the last glacial advance. Ridges similar to these have been observed in the country west of Hudson Bay by Bird (1953), who called them "drumlinoids".

Eskers are prominent topographic features in the east half of the area, where the drift cover is relatively thick. The eskers, which consist mostly of well-sorted sand, are as much as 40 feet wide and 20 feet high, but the average cross-sectional dimensions are about one-half of these figures. Although locally sinuous and discontinuous, they are for the most part remarkably linear and persistent. The fact that their trend is almost perpendicular to the latest direction of ice-movement presents an interesting problem. Perhaps the front of the melting ice-sheet was lobate, and sub-glacial drainage towards a salient gave rise to the anomalous eskers. The eskers could also have formed as deposits in fissures which opened parallel to the front of the retreating ice-sheet. Esker-like ridges formed in this way, called "ice-crack moraine", have been described by Sproule (1939). Undoubtedly the ice-sheet was in a stagnant condition as it melted, at least in the area where the anomalous eskers exist.

GENERAL GEOLOGY

General Statement

A major fault separates rocks of the Knob Lake group, in the southwestern part of the area, from rocks of the Murdock and Doublet groups, which underlie the central part of the area (see map in pocket). Its southeastern extension has not yet been investigated, but Frarey

(1952) has mapped this fault for a distance of more than 50 miles to the northwest. In accordance with previous mapping (Frarey, 1952), the Murdock and Doublet groups are placed stratigraphically above the Knob Lake group (Table 1).

Folds in the Knob Lake group plunge steeply to the northwest, in contrast to folds in the Murdock and Doublet groups which plunge gently to the southeast. Rocks of the Knob Lake, Murdock, and Doublet groups, together with the abundant conformable sills with which they are closely associated, represent the main rock-types of the Labrador Trough.

The eastern one-third of the area is underlain by gneisses and amphibolites which are separated from rocks of the Labrador Trough by a major fault zone.

All sedimentary rocks have been metamorphosed to some extent. Sandstones have been transformed to metaquartzites, siltstones to argillites, and shales to slates and phyllites. However, primary structures and textures have been preserved in many places, and, in this thesis, sedimentary names are used for rocks which exhibit little evidence of metamorphism.

KNOB LAKE GROUP

Harrison (1952) compiled the type section of the Knob Lake group on the basis of detailed study at Knob Lake, 35 miles west of Marion Lake. In Table 2, it is compared with the section exposed in the Marion Lake area. Sokoman, Ruth, and Fleming formations are missing in the Marion Lake area, and two formations, not present in the type section, have been mapped.

TABLE 1

Table of Formations

Era	Group	Formation	Lithology	Approximate Thickness
Cenozoic			Unconsolidated boulder till, sand, gravel	0-100'
		UNCONFORMITY		
Proterozoic			Basic dike rocks Gabbro, serpentinite, diorite, quartz diorite	
	Doublet		Meta-basalt; minor tuff and volcanic breccia	10,000'
			Slate, argillite, quartzite	3,000'
	Murdock		Chlorite schist, massive basic igneous rocks; minor sedimentary rocks	3,000'
		- - - Unconformity ?- - -		
	Knob Lake	Menihek	Slate, shale; minor quartzite, argillite	5,000'?
		Wishart	Sandstone, conglomerate, siltstone	300'
		- - - Unconformity - - -		
		Denault	Dolomite	3,000'
		Keco	Meta-basalt	2,500'
		Marion Lake	Sandstone, siltstone, slate, conglomerate	4,000'
		- - - Unconformity ?- - -		
		Attikamagen	Slate, phyllite	5,000'?
		FAULT CONTACT		
Archean ?			Gneiss, amphibolite, migmatite; minor schist	

TABLE 2

Stratigraphy of the Knob Lake Group

TYPE SECTION*

Formation	Lithology
Menihek Sokoman Ruth	Gray to black slate. Iron formation with slaty members. Black slate.
Wishart Fleming	Quartzite, arkose. Massive chert, chert breccia, chert conglomerate, quartzite.
Denault	Buff to gray weathering dolomite.
Attikamagen	Black, green, and gray slates; some granular chert.

*Modified after Harrison, 1952.

MARION LAKE**

Formation	Lithology
Menihek	Gray to black slate. Rusty weathering slate. Black slate.
Wishart	Quartzite Chert conglomerate. Cherty siltstone, quartzite.
Denault	Buff to gray weathering dolomite.
Keco	Metabasalt.
Marion Lake	Pink to white quartzite. Gray, green, and buff siltstone; conglomerate.
Attikamagen	Black to gray slate and phyllite; minor siliceous argillite.

**This report.

Ferruginous slates, which crop out on the north shore of Marion Lake, are possibly equivalent to the Sokoman and Ruth formations. However, because no breaks in the record of sedimentation were noted in the field, the rocks are here included in the Menihek formation.

Absence of the Fleming formation in the map-area may be due to erosion or local non-deposition. Harrison (1952, p. 9) notes that the Fleming in the Knob Lake area is thin and discontinuous. Chert fragments, which are abundant in the Wishart of the Marion Lake area, may have been derived from the Fleming.

A sequence of sedimentary rocks, herein named the Marion Lake formation, and a sequence of volcanic rocks, herein named the Keco formation, occur between the Attikamagen and Denault formations in the Marion Lake area. Because the lithology of the Marion Lake formation is quite similar to that of the Wishart, equivalents of the Marion Lake formation may have elsewhere been mapped as Wishart. Volcanic rocks closely associated with formations of the Knob Lake group have been termed "Nimish volcanics" by geologists of the Iron Ore Company of Canada. Perhaps the Keco formation has Nimish equivalents.

Attikamagen Formation

The Attikamagen formation is represented by widely scattered outcrops south and southeast of Marion Lake. Correlation with the type Attikamagen (Harrison, 1952) is based on lithologic similarity and stratigraphic position (below the Denault). The base of the formation is not exposed in the area, but steep dips and areal extent indicate a possible thickness of more than 5,000 feet. Attitudes of the strata are not consistent, however, and reversals in dip have been observed in outcrops

less than one hundred feet apart. This strongly suggests that the thickness of the formation has been exaggerated by structural repetition.

Description

Finely laminated, exceptionally fissile, rusty weathering, gray to black slates and silvery gray phyllites are the dominant rock types of the Attikamagen; a few interbeds of light-gray to bluish-gray siliceous argillite are exposed in the upper part of the formation. A typical Attikamagen outcrop is shown in Plate 2A.

Bedding consists of alternate quartz-rich and mica-rich laminations, commonly less than 0.5 mm thick, which form tabular units of relatively uniform thickness. The micaceous layers are finer grained and more carbonaceous than the siliceous layers. Small-scale cross bedding can commonly be seen in the silty layers. Primary lamination and structure of the phyllites are obscured by flow and fracture cleavage (Plate 2B).

Grain size for most Attikamagen rocks ranges from 0.005 to 0.1 mm. Some minerals in the phyllites can be identified by petrographic methods, but accurate modal analyses cannot be obtained because of the fineness of grain. The average mineral composition, based on estimates for several thin sections, is approximately: 50% quartz, 25% chlorite, 15% muscovite, and 10% albite. In some phyllites, biotite takes the place of muscovite and chlorite. Although much of the quartz and albite is recrystallized, dusty outlines within a few grains indicate detrital origin.

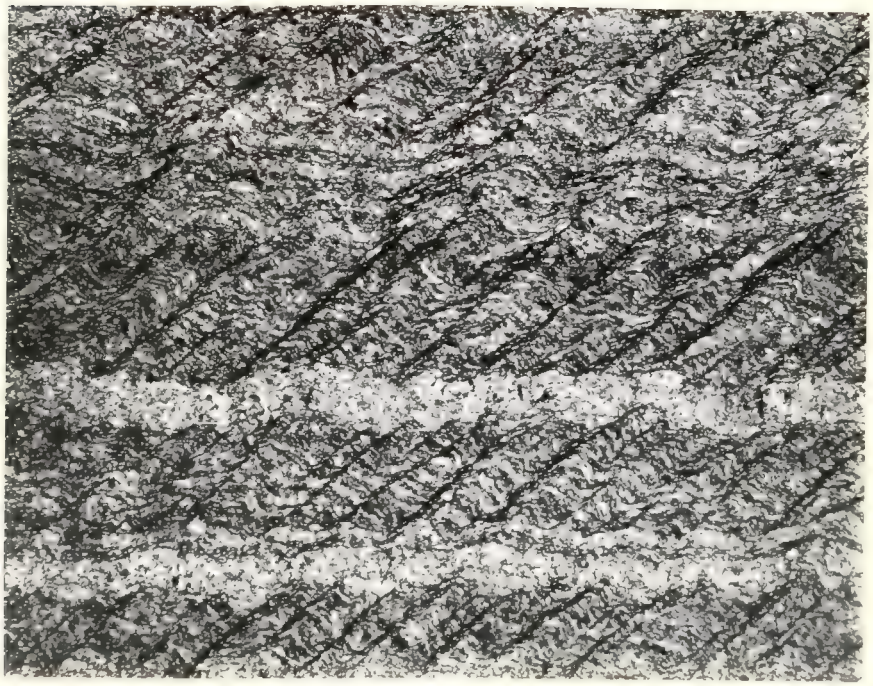
Because the black slates are extremely fine-grained and are commonly clouded with carbon, their mineralogy cannot be satisfactorily

PLATE 2

Attikamagen slate and phyllite

A. Typical outcrop of Attikamagen slate showing well-developed bedding cleavage. White object near center of photograph is two inches wide.

B. Thin section of phyllite. Bedding, consisting of alternate quartz- and mica-rich layers, is crenulated by inclined fracture cleavage. Ordinary light, X25.



determined by petrographic methods. Chemical analyses and x-ray studies are required to determine the mineral composition. The chemical composition of a representative slate sample from the Marion Lake area, and the compositions of two Attikamagen samples from the type locality (Gross, 1951), are shown in Table 3.

The three samples have notably high silica and low lime contents, and in contrast to normal pelitic rocks, they contain no carbonate minerals (indicated by the lack of CO_2). The sample from the Marion Lake area differs from the other two in having a lower potash content, a higher magnesia content, and a much higher ferrous to ferric iron ratio. X-ray analysis of sample 58-93A revealed that the major minerals are quartz, chlorite, muscovite, and albite. The normative composition of the sample is:

Quartz	47.0	Albite	12.7
Chlorite	25.7	Sphene	0.9
Muscovite	12.2	Carbon	0.2

Origin and Depositional Environment

Absence of coarse clastics in the Attikamagen suggests deposition far from shore, but may also be explained by near-shore sedimentation adjacent to a low-lying land mass. The delicate lamination indicates relatively quiet water deposition, but common occurrence of cross-bedding demonstrates at least intermittent current or wave action, suggesting that depth of water was not great. Rich says of similar finely laminated shales (1951, p. 2022): "(They) must have been deposited in quiet water free from wave or current action," but he notes (1951, p. 2039) that: "the depth of water need not have been more than a few hundred feet."

TABLE 3

Chemical Composition of Attikamagen Samples

Specimen No.	58-93A	L2	L3
SiO_2	68.5	64.43	68.72
TiO_2	0.4	0.66	0.53
Al_2O_3	14.3	18.39	14.69
Fe_2O_3	0.7	2.71	2.83
FeO	5.91	2.56	1.67
MnO	---	---	---
MgO	2.1	1.69	1.40
CaO	0.2	---	---
Na_2O	1.5	0.80	2.02
K_2O	1.5	4.92	4.35
H_2O +	3.38	3.75	3.58
H_2O -		0.24	0.42
P_2O_5	---	0.03	T
CO_2	---	---	---
SO_3			
C	1.46	---	---
Total	99.9	100.18	100.20

58-93A Black slate, Attikamagen formation, Marion Lake area.
Analyst - K. Hoops, Geol. Surv. Canada

L2, L3 Slate, Attikamagen formation, Knob Lake area.
Analyst - W. H. Herdsman, Geol. Surv. Canada

Carbon in the Attikamagen suggests contribution of organic material, but because fossils are absent, the source of carbon is unknown. Excellent preservation of lamination indicates the absence of bottom-dwelling organisms (Moore and Scruton, 1957, p. 2743), and probably, therefore, pelagic organisms were the contributing life forms. The lack of carbon in the two samples from the Knob Lake area may reflect unequal density of biota within the basin, but could also be the result of local variations in oxidation potential.

Twenhofel (1939) reviewed the black shale problem and described the environmental conditions (1939, p. 1181) as due to: "... poor circulation and ventilation of water, limited entrance of oxygen connected with the limitation of circulation and accumulation of toxic products." Some circulation must have occurred in the Attikamagen, as indicated by the scattered cross-bedding, but it obviously was not sufficient to bring about complete oxidation of organic material. Pettijohn (1957, p. 625-626), in a discussion of the anaerobic environment necessary for deposition of carbonaceous shales, states: "It seems most probable that some impediment to convection is necessary and the best hindrance is a density stratification related to salinity differences. Such is readily obtained in a barred basin." The main requirement of the barred basin environment is an offshore barrier (Fleming and Revelle, 1939), but the existence of such control cannot be demonstrated for the Attikamagen.

Marion Lake Formation

The name, Marion Lake formation, is here proposed for the sequence of sandstones, siltstones, slates, and conglomerates which overlies the Attikamagen. Table 4 shows the stratigraphy of the

TABLE 4

Stratigraphy of the Marion Lake Formation

Unit	Lithology	Approximate Thickness
8	Rhythmically bedded, buff, pale green and dove gray siltstone; lenses of conglomerate and sandstone.	600'
	- - - Basic Sill - - -	
7	Rhythmically bedded, pale green and buff siltstone.	1000'
6	Massive, cross-bedded, cream, pink, and pale green sandstone.	200'
5	Intercalated rusty- to gray-weathering dark slate and gray, very fine-grained sandstone with small-scale cross-bedding.	600'
4	Pale green to buff rhythmically bedded siltstone.	100'
3	Massive pink sandstone.	100'
	- - - Basic Sill - - -	
2	Rhythmically bedded gray and white siltstone; laminated pink siltstone.	1200'
1	Cross-bedded pink sandstone; massive, pale yellow sandstone; interbedded conglomerate.	200'
	Approximate total thickness	4000'
	Rhythmically bedded siltstone	2800'
	Sandstone	1100'
	Black slate	60'
	Conglomerate	40'

formation. Thicknesses were scaled from aerial photographs and corrected for dips measured in the field. It is assumed that the three bands of sedimentary rock, now separated by basic sills, were contiguous prior to intrusion. On the basis of lithology, the formation has been divided into eight units. Because the base of the lowermost unit lies beneath the southwest arm of Marion Lake, the nature of the contact with the Attikamagen is unknown. However, slate and quartzite pebbles in conglomerate beds near the base of the Marion Lake formation suggest an unconformity; chert beds in the Attikamagen have been reported by Harrison (1952, p. 8), and quartzites ("Seward grits") stratigraphically below the Attikamagen are reported by geologists of the Iron Ore Company of Canada (Glenn Hogg, personal communication, 1958).

Siltstones make up about 70 percent of the Marion Lake formation (Units 2, 4, 7, and 8), sandstones less than 30 percent, (Units 1, 3, 5, and 6), conglomerates about 1 percent (30 feet of Unit 1; 10 feet of Unit 8), and black slates less than 1 percent (interbeds in Units 5 and 8).

Sandstone

Description

The sandstones of Units 1, 3, and 6 are tough, massive, medium- to fine-grained, ridge-forming rocks which are pink, tan cream, or pale-yellow on fresh and weathered surfaces. Bedding, where visible, is marked by streaks of iron oxide and heavy minerals, and ranges from 1 to 5 mm in thickness. Tabular units showing planar cross-lamination occur in a few outcrops of Unit 1. More abundant is cross-lamination of the festoon type, consisting of wedge-shaped, cross-bedded units, 3 to 8

inches thick and 1 to 3 feet long, (Plate 3A). Inclinations of both types of cross-bedding range from 10° to 27°.

The attitudes of 14 planar cross-beds were measured in sandstone outcrops, west of Marion Lake. The attitudes of 10 cross-beds were also measured in outcrops of Wishart sandstone at Squaw Lake, 40 miles west of the map-area. The data, corrected for bedding tilt by the stereonet method (Haff, 1938), are summarized and compared in Fig. 2. Both sets of data indicate predominant sedimentary transport from the west.

Most sandstone outcrops show thin, even, planar bedding. Commonly associated with the normal bedding, however, are tabular units, 1 to 3 inches thick, which contain small, concave-upward ripple-like structures. In longitudinal section they appear as symmetrical, curved traces with amplitudes less than 0.5 inches and wave-lengths of 1 to 3 inches. The "ripples" are commonly truncated, as a result of overlap, in a fashion suggesting scour and fill.

Unit 5 consists of intercalated black slates and very fine-grained, rusty-weathering, light gray sandstones. The slate layers, which are 1 to 5 mm thick, alternate rhythmically with sandstone layers 1 to 4 cm thick. Many sandstone layers are cross-bedded and tend to pinch and swell laterally. Contacts between the sandstone and slate beds are sharp; gradation in composition or grain size from bottom to top of individual beds is notably absent.

The sandstones contain 46 to 92 percent quartz. Most quartz grains exhibit clastic boundaries, but only a few have overgrowth rims. The grains are subangular to well-rounded and are highly spherical (Plate 4). Most of the quartz is strained, but very few grains are fractured. Clear, sharply-twinned detrital plagioclase (mostly oligoclase),

Fig. 2

Cross-bedding dip azimuths for two localities in the Labrador Trough. The data are corrected for bedding tilt and plotted on rose diagrams to show percent frequency distribution in 30° class intervals. The circular traces define 10 percent increments.

A. Rose diagram based on 14 cross-beds in Unit 1 of the Marion Lake formation. Readings taken on the southwest shore of Marion Lake.

B. Rose diagram based on 10 cross-beds in an outcrop of Wishart quartzite. Readings taken at Squaw Lake, 35 miles west of the Marion Lake area.

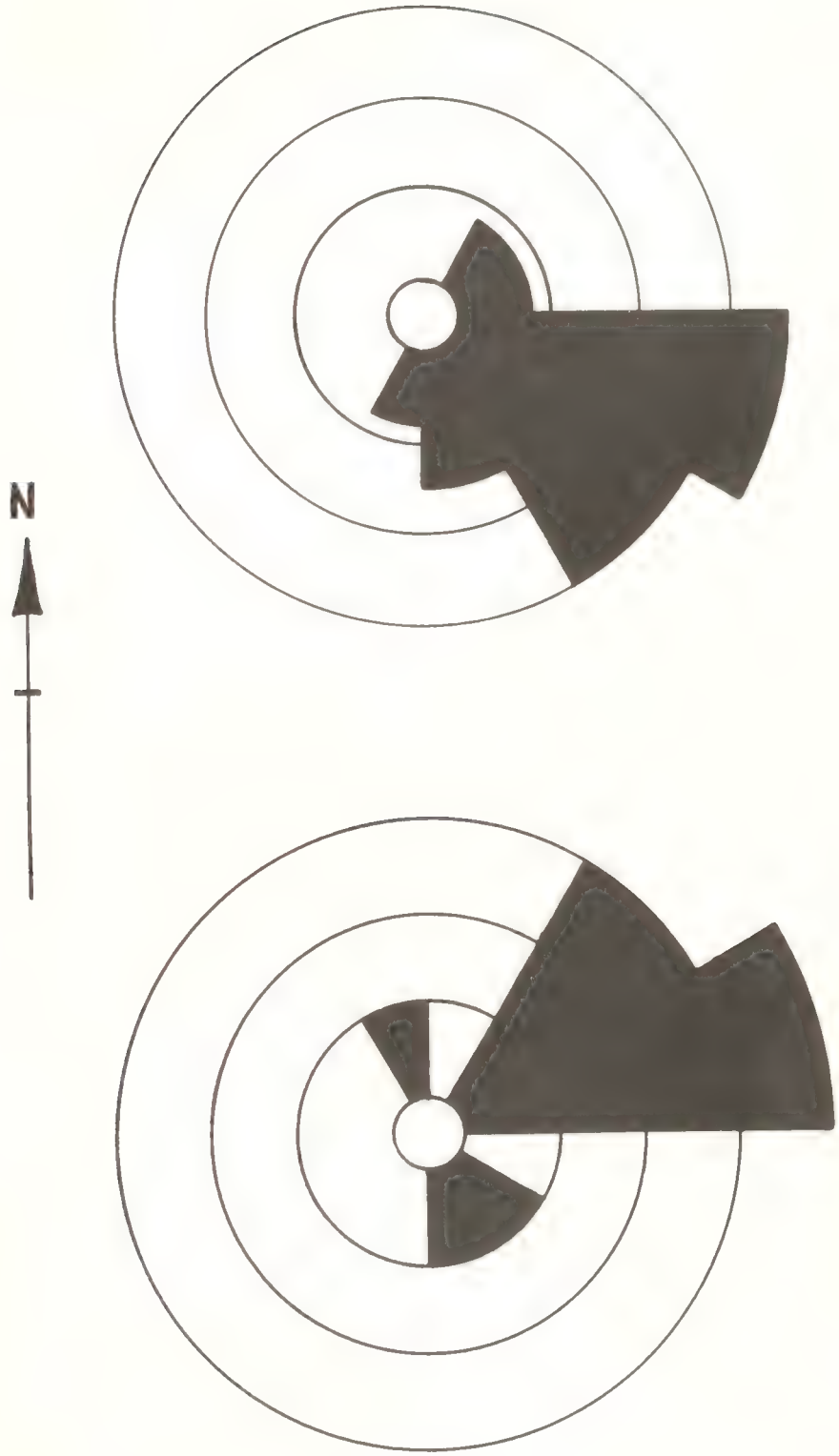

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PLATE 3

Outcrops of Marion Lake sandstone

A. Arkose of Unit 1 showing festoon cross-beds. Field seen is 2 feet wide.

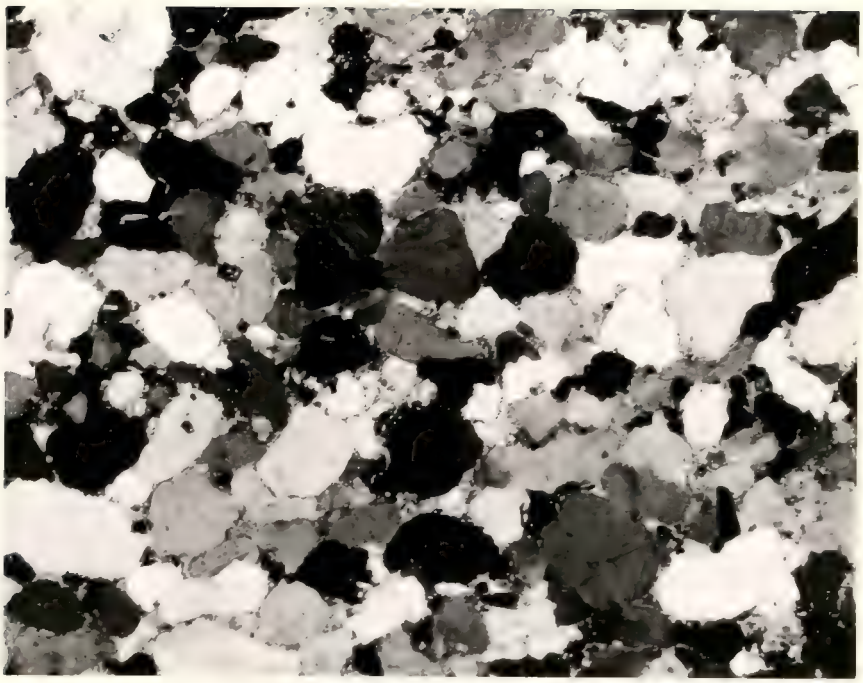
B. Jointed, partly frost-heaved exposure of sandstone. Scale indicated by hammer.

PLATE 4

Photomicrographs of Marion Lake sandstone

A. Orthoquartzite consisting of well-sorted, well-rounded, fairly spherical grains of quartz. Note microstylolitic interpenetration of grains. Crossed nicols, X40.

B. Arkose containing abundant quartz and fresh plagioclase grains. Compare sorting and rounding with (A). Crossed nicols, X30.



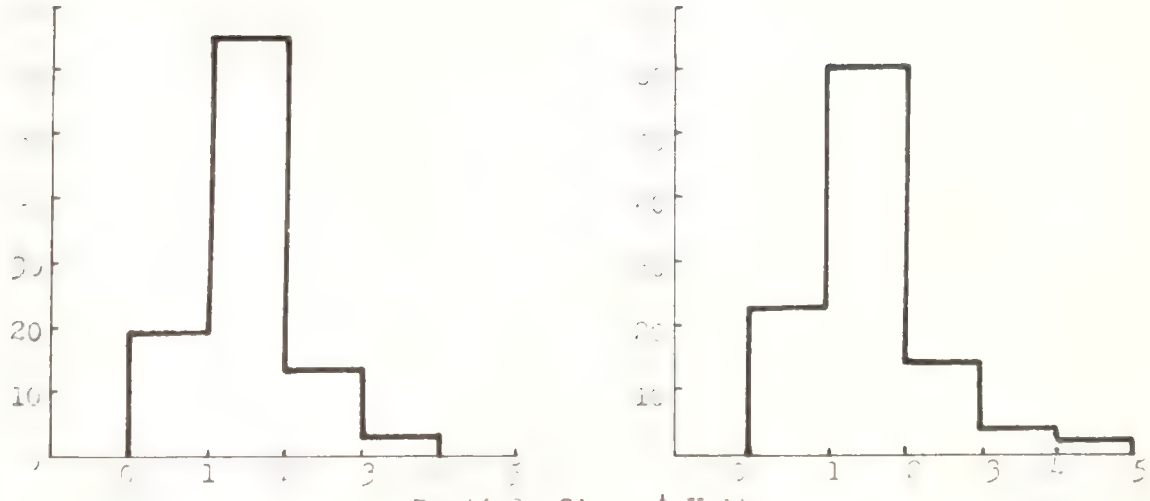
is present in all but one sandstone thin section, and small amounts of untwinned metamorphic albite are present in all of them. Potash-feldspar is relatively abundant in one specimen, but scarce in the others. Most of it is clear microcline, easily recognized by the grid twinning. The feldspars have the same roundness, sphericity, and size distribution as the quartz. Some quartz and feldspar grains are marginally replaced by fine-grained micas, and some have microstylolitic boundaries in common with adjacent grains. Most specimens of the Marion Lake sandstone contain a few rock fragments which are similar in size and shape to the quartz and feldspar grains. Chert is the most abundant rock fragment, but siltstone, quartzite and slate are commonly present. Size distributions for several sandstones were calculated according to the method outlined in Appendix A. Modal size classes are in the range of fine- to medium-grained sand, and more than 90 percent of each sample is contained by three Udden size classes (Fig. 3). The sandstones are well-sorted (Trask, 1932, p. 71).

Silt- and clay-sized flakes of muscovite are major constituents of the matrix. X-ray analysis indicates that the muscovite has a 2M structure (Yoder and Eugster, 1955). It is intimately intergrown with chert and, in places, chalcedony. Biotite and chlorite, which are the only other quantitatively important matrix minerals, are relatively scarce in most of the thin sections.

Detrital tourmaline and zircon are common accessory minerals. The tourmaline and zircon grains are less than 0.1 mm in diameter and are well-rounded. Some of them exhibit euhedral overgrowths. The opaque minerals include magnetite, ilmenite, leucoxene, and hematite. Pyrite is abundant in Specimen 57-50B, but is absent in the other sandstones.

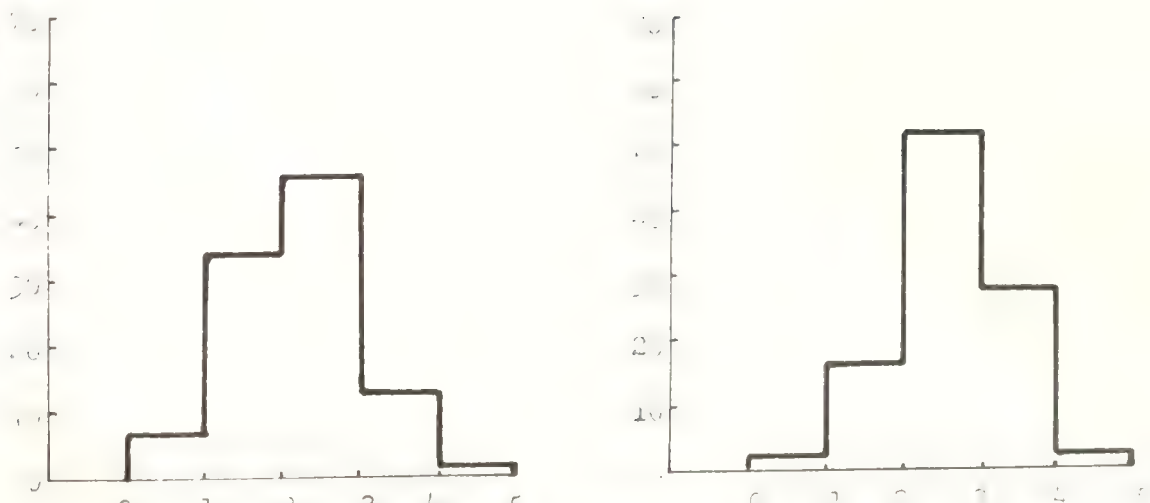
Fig. 3

Size histograms for four Marion Lake sandstones, showing the estimated volume percent distribution of detrital grains.



57-5Ln

57-13Gn



57-13Gn

57-13Gn

The pyrite is probably authigenic, because it occurs in euhedral grains and is not concentrated along the bedding.

Modal compositions of five sandstone specimens are shown in Table 5. According to Pettijohn's classification (1957, p. 291), the predominant rock types are feldspathic sandstones, but one specimen, which is representative of Unit 6, is an orthoquartzite. The chemical composition of a representative orthoquartzite from the Marion Lake formation is shown in Table 6.

TABLE 6

Chemical Composition of a Marion Lake Orthoquartzite*

SiO_2	95.4	CaO	0.0
TiO_2	0.0	Na_2O	0.1
Al_2O_3	2.9	K_2O	0.9
Fe_2O_3	0.2	H_2O	0.52
FeO	0.16	P_2O_5	0.0
MnO	0.0	CO_2	0.00
MgO	0.4	Total	100.6

*Sample 57-51A Orthoquartzite, Unit 6, Marion Lake formation.
Analyst: G. Bender, Geol. Surv. Canada

The orthoquartzite, which lacks lime and has high alkali and alumina contents, is similar in chemical composition to the Lorrain quartzite (Collins, 1925, p. 68). The Marion Lake sample has the following normative composition:

Quartz	89.5	Muscovite	7.3
Albite	0.8	Chlorite	1.4

TABLE 5

Modal Composition of some Marion Lake Sandstones

(Volume percent, based on 800 points for each specimen)

Specimen No.	Unit	Quartz	Plagioclase	Microcline, Orthoclase	Rock Fragments	Muscovite	Biotite, Chlorite	Opaques	Zircon	Tourmaline	Silica Cement	Average Roundness	Average Sphericity	Median Diameter, mm	S_o, Sorting Coefficient	Classification (Pettijohn, 1957)
57-51A	6	91.3	*	—	—	7.7	1.0	*	*	—	*	0.9	0.8	0.58	1.28	Orthoquartzite
57-130A	1	61.8	9.5	5.2	1.8	12.8	*	0.8	*	*	8.1	0.2	0.5	0.60	1.42	Subarkose
57-132A	1	71.0	16.0	*	1.7	8.0	*	0.8	*	*	2.5	0.6	0.7	0.32	1.38	Arkose
57-133A	1	62.5	12.3	*	*	13.1	*	0.9	0.3	0.7	10.2	0.3	0.4	0.22	1.35	Subarkose
57-50B	5	46.4	5.1	*	*	9.7	5.1	7.5	*	—	26.2	0.2	0.6	0.11	1.45	Subarkose

* Less than 0.2 percent.

Note that the modal composition (Table 5) for the orthoquartzite closely corresponds to the normative composition (p. 20).

Origin and Depositional Environment

Size distributions of the well-sorted Marion Lake sands (Fig. 3) are similar to those of modern sands deposited in shallow water, fluviatile, and aeolian environments (Udden, 1914, p. 703; p. 714-719, and Wentworth, 1932, p. 36-75).

The abundant festoon cross-beds are indicative of scour and fill action by shifting currents. Knight (1930) attributes the formation of festoon cross-bedding in the Casper and Tensleep formations to "oscillating currents in comparatively shallow water," but McKee (1953, p. 27, 31, and Fig. 22) has described festoon cross-bedding in alluvial fans and aeolian sand dunes. Bagnold (1941, p. 201) and McKee (1953, p. 27) have noted that aeolian cross-beds commonly have inclinations ranging from 30° to 34°. Inclinations of cross-beds measured in Marion Lake sandstones are all less than 27°, suggesting subaqueous deposition.

The Marion Lake sandstones which exhibit fine, planar lamination closely resemble sand deposits of modern beaches. Thompson (1937, p. 727) describes California beach sand in which single laminations can be traced 100 feet or more. The common occurrence of ripple-like cross-bedding in the sandstones showing planar bedding is noteworthy, because Brinkman and Hulseman (Cited by Pelletier, 1957, p. 68) have observed such cross-bedding in modern littoral environments.

The sandstone-slate interbeds of Unit 5 are similar to sand-shale associations now being deposited on tidal flats of the Wadden Sea. The Wadden Sea sediments show the following features (Van Straaten, 1951):

1. Alternation of sand and clay layers
2. Lack of graded bedding
3. Abundance of cross-bedding in the sandstone layers
4. Lateral irregularity in thickness of individual beds
5. Sands are very fine-grained

The Unit 5 rocks exhibit all these features, and the scale of the interbedding (beds less than 4 cm thick) approximates the scale of the tidal flat deposits. Potter and Glass (1958) describe similar siltstone-shale interbedding in a sequence of Pennsylvanian age which is interpreted as forming, in part, on a "shallow, marginal, marine shelf." These workers note similarity of Texas Gulf coast sediments, Wadden Sea sediments and the rocks which they describe.

Most of the Marion Lake sandstones are somewhat feldspathic. Krynine (1943) pointed out that the presence of abundant feldspar can reflect either high relief of the source area or a rigorous climate. The poor rounding exhibited by the feldspathic sandstones indicates a short abrasional history, and therefore source-area uplift is the favoured explanation for the feldspar influx. As demonstrated by cross-bed data, direction of sedimentary transport was from the west. The quartz- and feldspar-rich crystalline rocks which flank the west side of the Labrador Trough are probable source rocks. Uplift along the west side of the Trough may therefore have initiated the periods of arkose deposition.

Conglomerate

Description

Layers and lenses of conglomerate, less than 15 feet thick, are conformably interbedded with sandstones in the lower parts of Units 1 and

8. The conglomerates are gray- to tan-weathering, greenish-gray to buff rocks which are massive and extremely well-indurated. Pebbles and granules of quartzite and chert compose 25 to 35 percent of the conglomerates (Plate 5). A few slate fragments are present in some specimens. Most of the siliceous rock fragments are well-rounded and equidimensional, but the slate fragments have distorted lenticular shapes. Frameworks are disrupted; the rock fragments are enclosed by a matrix similar in composition to the associated arkosic sandstones.

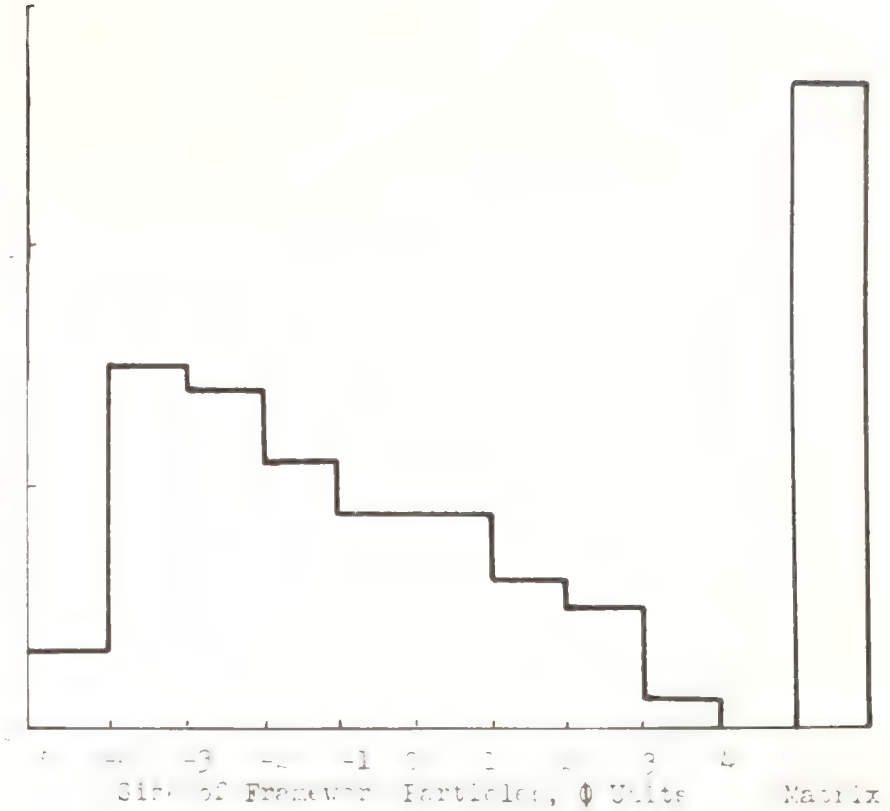
Origin and Depositional Environment

Estimated size frequency distributions for two Marion Lake conglomerates are shown in Fig. 4. The distributions resemble the distributions of modern alluvial gravels (Wentworth, 1932, p. 32-37) and flood gravels (Krumbein, 1940, 1942). The poor sorting reflects a short history of transport. Some of the pebbles are well rounded, but studies of recent gravels (Plumley, 1948) have shown that pebbles become well rounded by stream transport within relatively short distances. Beach gravels are commonly well rounded, but, unlike the Marion Lake conglomerates, show good sorting. Glacial till is characterized by a wide range of size grades (Krumbein, 1933), but because the Marion Lake conglomerates form lenses and layers intercalated with well-sorted sandstones, glacial origin is unlikely. Lacking are cobbles, boulders, facetted and striated pebbles, glacial pavements, and other evidence of glacial origin. The conglomerates are perhaps best interpreted as rapidly deposited alluvial gravels which invaded the environment of sand deposition.

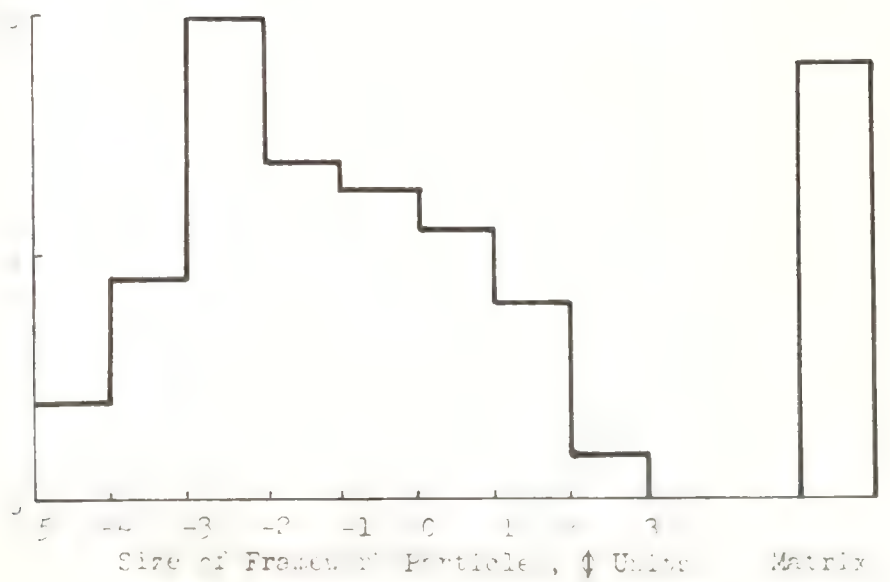
The Attikamagen formation is the only known source for the chert and slate fragments in the conglomerates, and the Seward formation is the

Fig. 4

Size histograms for two Marion Lake conglomerates, showing the estimated volume percent distribution.



57-136C



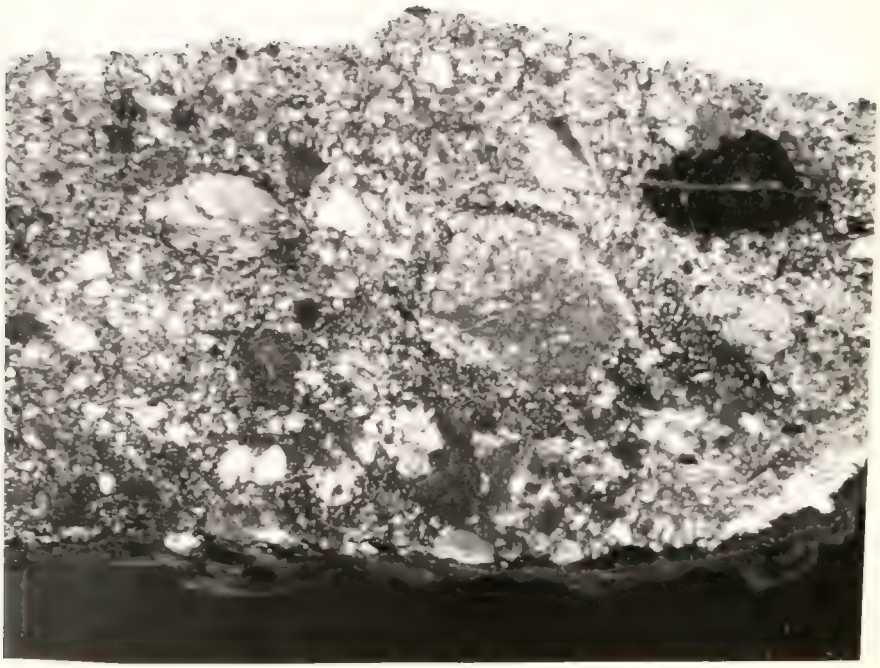
57-152A

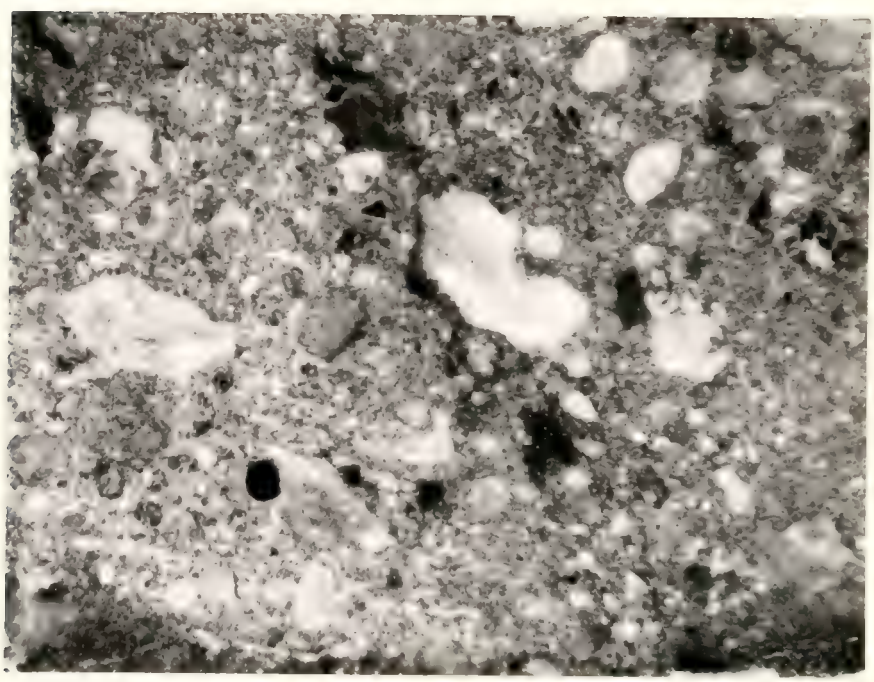
PLATE 5

Marion Lake conglomerate

A. Polished surface of Unit 1 conglomerate showing abundant pebbles of quartzite (white and light gray) and chert (dark gray and black), X1.5.

B. Polished surface of Unit 8 conglomerate. Pebbles of quartzite and chert (dark gray), X1.5.





probable source of the quartzite fragments. An unconformity at the base of the Marion Lake formation is therefore inferred, although the contact is not exposed in the map-area. Because some conglomerates occur in the upper part of the Marion Lake formation, the source rocks must have either occupied an area of non-deposition during Marion Lake times, or have been re-exposed as a result of local uplift and erosion.

Siltstone

Description

Rhythmically bedded siltstones compose approximately 70 percent of the Marion Lake formation. The rhythmic sedimentation unit, or couplet, consists of a basal silt-rich siliceous layer and an upper micaceous layer in which clay-size particles predominate. Boundaries between couplets are sharp, but contacts between the coarse-grained siliceous and fine-grained micaceous layers of individual couplets are generally gradational. The gradation in size and composition commonly cannot be seen on fresh or weathered surfaces of hand specimens, but is easily recognized in thin section and on polished surfaces cut perpendicular to the bedding. In the upper portion of Unit 2, however, where the silty layers have weathered cream to white and the finer-grained layers have weathered drab greenish gray to black, the grading is quite prominent in outcrops (Plate 6A).

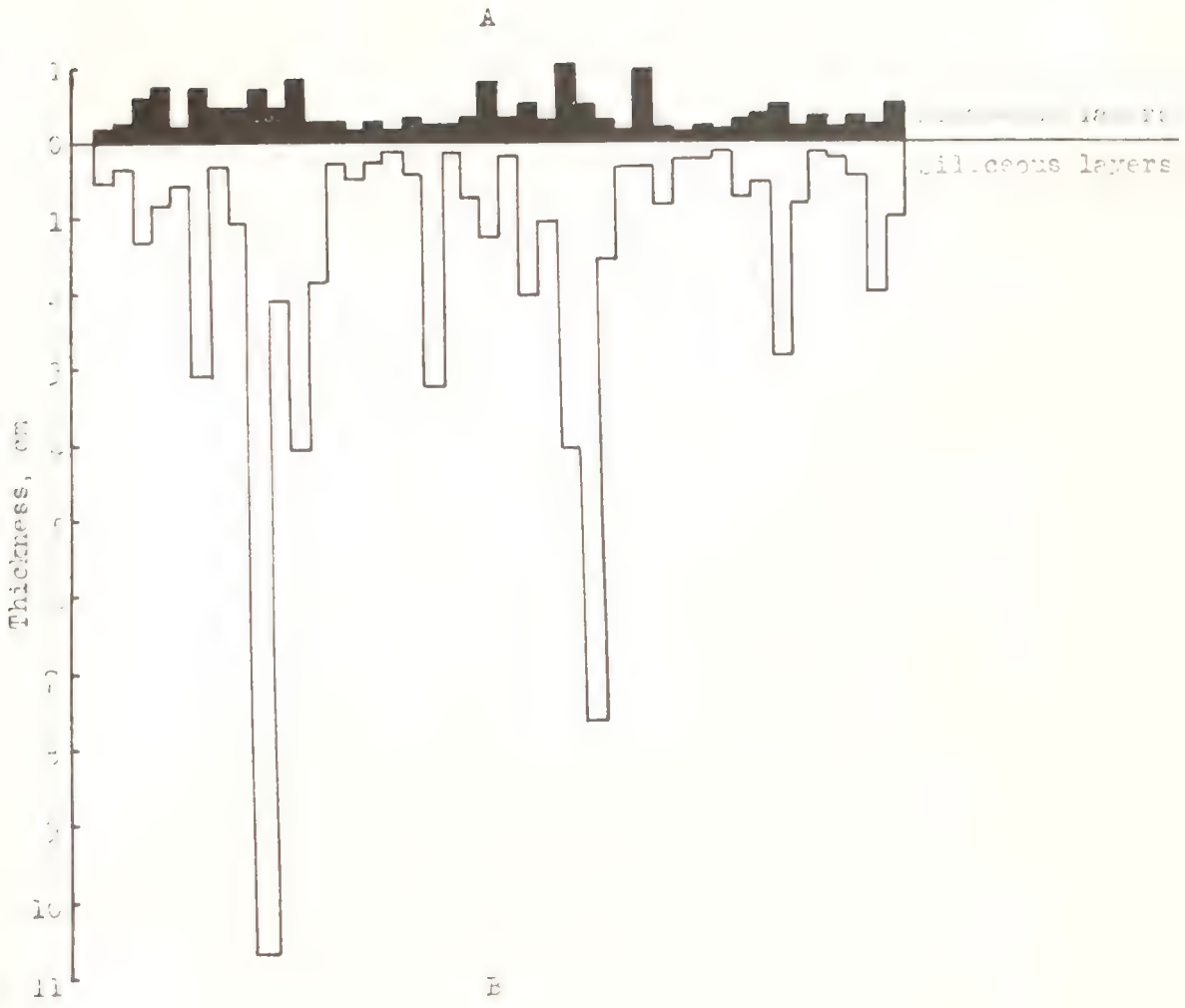
The silt-rich layers are 0.5 to 10 cm thick and all clay-rich layers are less than 1 cm thick. Variations in the thicknesses of 42 successive couplets are illustrated in Fig. 5A. The total thicknesses of these beds were grouped in classes, and the cumulated frequencies in each class were plotted on logarithmic probability paper. The distribution is log normal (Fig. 5B).

Fig. 5

Thickness data for graded beds in Unit 2 of the Marion Lake formation.

A. Thicknesses of the siliceous and micaceous layers of 42 successive bipartite beds. Oldest couplet to the left.

B. Cumulated frequency distribution of the 42 total thicknesses, plotted on a logarithmic probability chart.


A
Thickness, cm
B

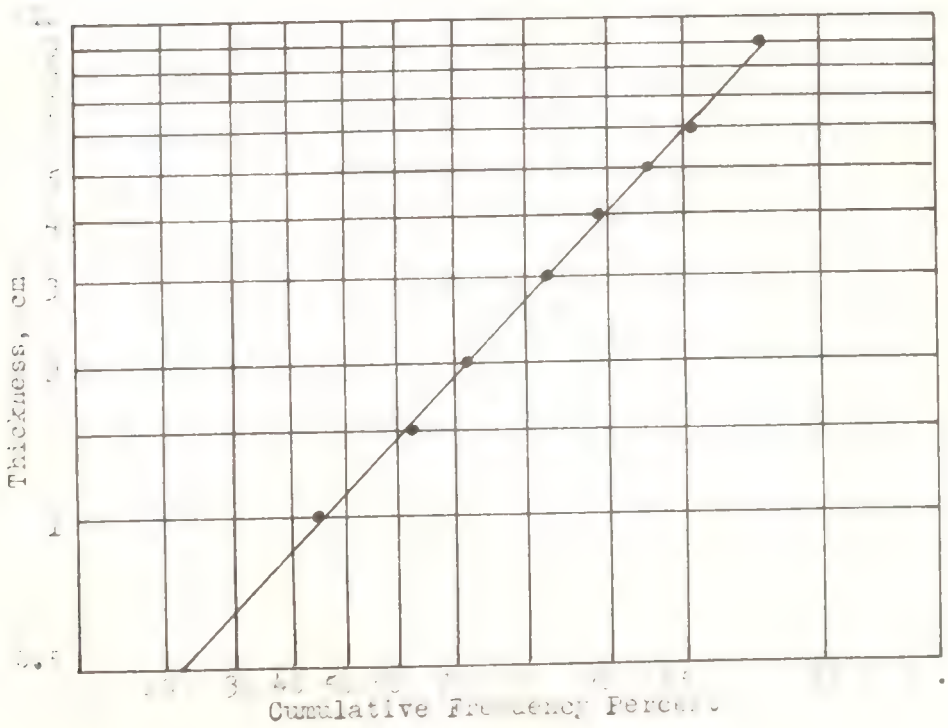
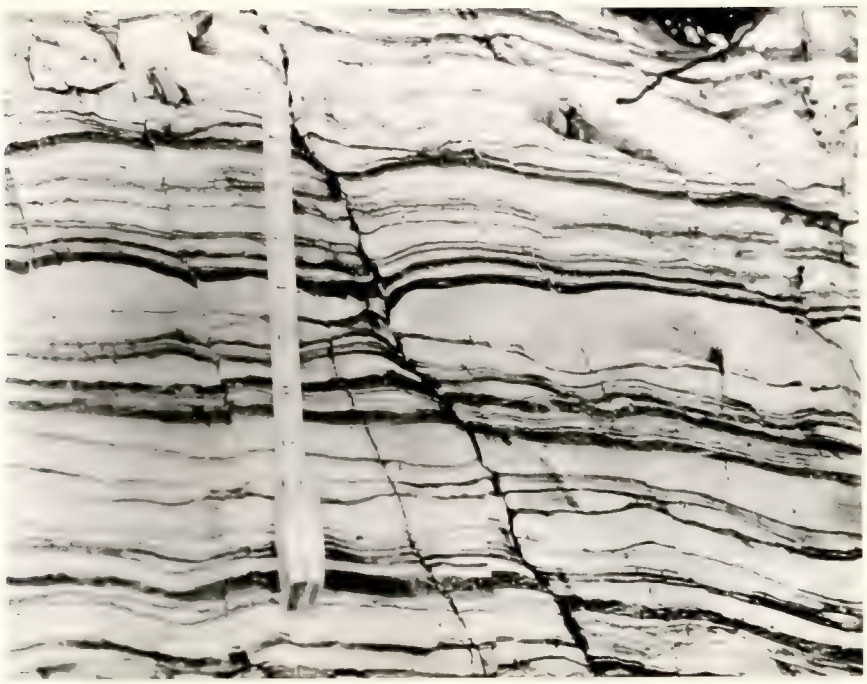

Thickness, cm

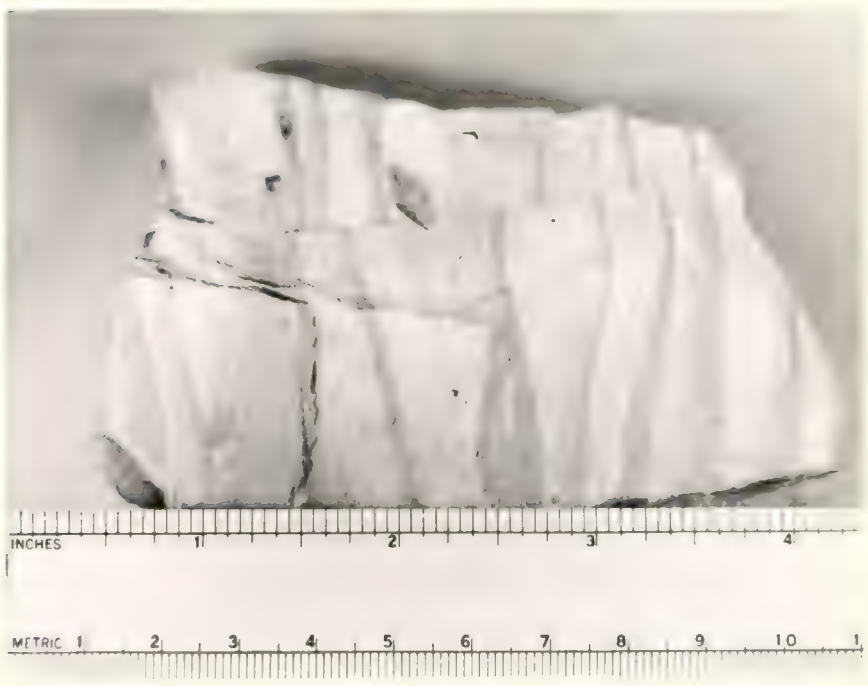
PLATE 6

Rhythmically bedded Marion Lake siltstone

A. Outcrop showing alternation of light silty layers and dark argillaceous layers.

B. Hand specimen showing graded bedding and fine lamination within the graded beds. Top of specimen is to the right.




INCHES
1
2
3
4
METRIC 1
2
3
4
5
6
7
8
9
10

Most of the graded beds are finely laminated (Plate 6B). The laminations, which are tabular and less than 1 mm thick, are commonly parallel to the boundaries of the bed in which they occur, but in some silty layers, they form planar cross-stratified units (Plate 7A). Many laminations are graded in the same way as are the beds in which they occur.

The largest quartz grains in the siltstones, which occur at the base of the beds, have maximum diameters of 0.05 to 0.3 mm, but the average grain size is less than 0.05 mm. Grain size diminishes towards the tops of the beds; few grains in the upper layers exceed 0.01 mm in diameter. The quartz grains are equant and subangular to angular, and most of them are slightly strained. Fresh, sharply-twinned albite is present in all the siltstones, and occurs in grains of the same shape and size as the quartz. Detrital outlines are visible in some larger quartz and plagioclase grains.

The micaceous minerals, muscovite and chlorite, are interstitial to the quartz and albite in the coarse-grained layers and are the predominant minerals in the fine-grained layers. A few flakes of muscovite are probably detrital, because they show a size gradation corresponding to the size gradation of silt particles. However, most of the muscovite and chlorite does not show this variation, and probably represent recrystallized clay minerals.

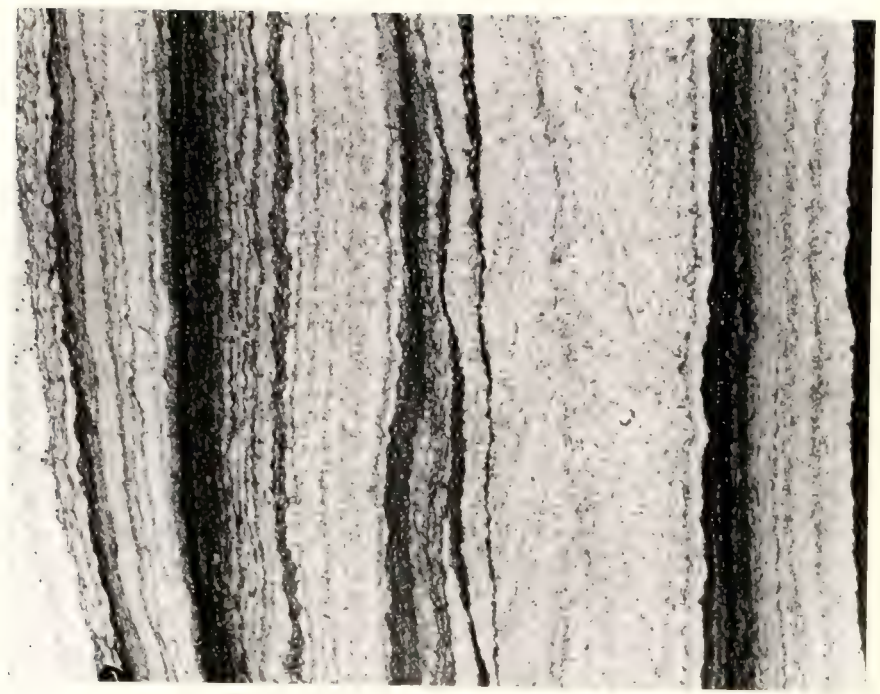
Limonite is common in the siltstones as patches, up to 1 mm in diameter, dispersed through the silty layers, and as surface stains along fractures and cleavage surfaces. The limonitic patches apparently represent weathered pyrite grains, because pyritic cores are present in some of the patches, and euhedral pyrite of probable authigenic origin is

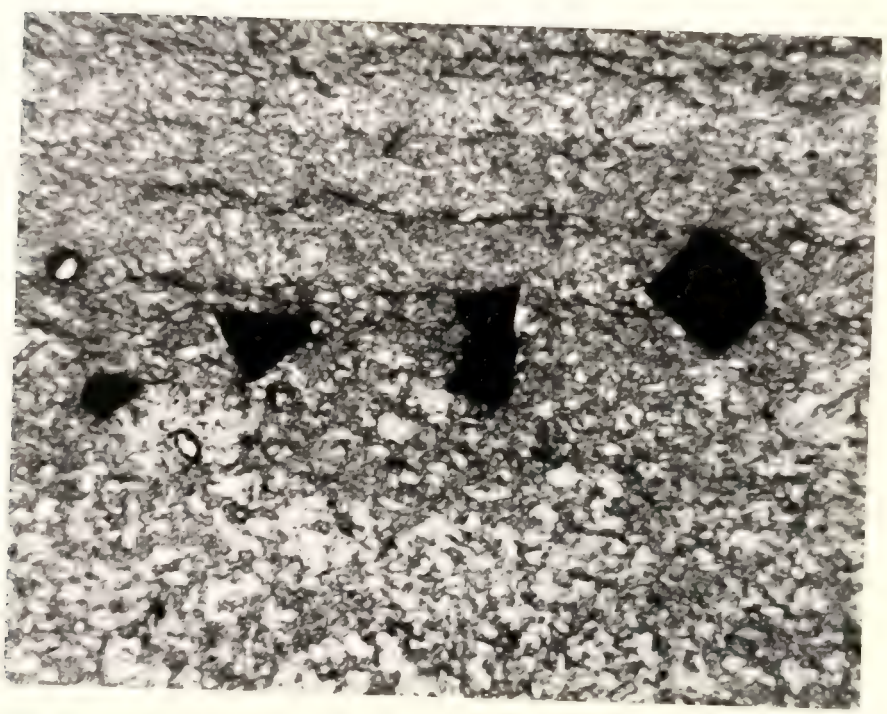
PLATE 7

Photomicrographs of Marion Lake siltstone

A. Thin section of rhythmically bedded siltstone. Note cross-stratification in thickest layer. Ordinary light, X10.

B. Authigenic pyrite (black) in siltstone. Note truncation of bedding by the euhedral pyrite grains, ordinary light, X50.





present in unweathered specimens of the siltstones (Plate 7B). Detrital zircons, less than 0.1 mm in diameter, were seen in all thin sections of siltstones, and euhedral rods of pleochroic bluish-green tourmaline, as much as 0.2 mm long, are abundant in some specimens. Most of the tourmaline crystals have rounded detrital cores. Carbon is disseminated through the dark, fine-grained layers of the Unit 2 siltstones. Apatite, magnetite, and epidote are visible in some siltstone thin sections.

Modal compositions of the silt-rich layers in the rhythmically bedded siltstones are shown in Table 7. Components of the micaceous layers cannot be readily distinguished. The chemical composition of a representative siltstone sample is shown in Table 8. The normative composition of the sample, calculated in terms of the minerals identified in thin section is:

Quartz	28.2	Sphene	0.4
Albite	23.2	Illmenite	0.9
Muscovite	35.4	Iron Oxides	1.7

Comparison of this data with the modal analyses of the coarse-grained layers, (Table 7), indicates that the irresolvable fine-grained layers, which compose about one-third of the rhythmically bedded siltstones, must consist almost entirely of muscovite and chlorite.

TABLE 7

Modal
Composition of Some Marion Lake Siltstones

(Modal Analyses based on 600 Counts)

Specimen No.	Unit	Quartz	Plagioclase	Muscovite	Chlorite	Carbonate	Opaques	Zircon	Tourmaline
57-131A	2	34	27	30	6	*	3	*	*
57-134A	2	45	23	12	16	---	4	*	*
57-50A	4	37	15	36	9	---	2	*	*
57-50C	4	38	17	32	10	*	3	*	0.6
57-53A	7	48	12	24	14	---	2	*	1.8
57-147A	8	41	22	23	11	2	1	*	*

*Less than 0.5 percent.

TABLE 8

Chemical Composition of a Rhythmically Bedded Siltstone* from the Marion Lake Formation

SiO_2	61.7	CaO	0.2
TiO_2	0.7	Na_2O	2.7
Al_2O_3	19.7	K_2O	4.1
Fe_2O_3	1.7	H_2O	2.89
FeO	2.80	P_2O_5	0.0
MnO	0.0	CO_2	0.04
MgO	2.0	Total	98.5

*Sample 57-131A.

Analyst: K. Hoops, Geol. Surv. Canada

Origin and Depositional Environment

The common occurrence of cross-bedding in the Marion Lake siltstones indicates some current action, but the excellent preservation of delicate laminations and graded bedding precludes the possibility of deposition above wave base. The environment of deposition must therefore have been one of relatively tranquil water. However, the siltstones are interbedded with sandstones and conglomerates which apparently were deposited in shallow water and subaerial environments. It is therefore unlikely that the siltstones were deposited at great depth, as this would require extensive upwarping and downwarping during a geologically short period. Deposition immediately below wave-base is therefore postulated for the Marion Lake siltstones, as this would necessitate the least tectonic activity at the site of deposition, or, in the case of eustatic change in sea level, no tectonic activity.

The graded beds of the Marion Lake siltstones resemble in many ways the varves of Pleistocene deposits (De Geer, 1912; Sauramo, 1923), but they do not show any evidence of glacial origin, such as association with tillites or presence of rafted pebbles. The graded beds may, however, have formed in the same way as glacial varves, in response to annual climatic change. The abundance of micrograded laminations within the graded beds precludes the possibility that the couplets represent a single event, as do graded beds of graywacke deposits (Kuenen and Migliorini, 1950). Any hypothesis of origin must account for the sub-graded laminations within the graded couplets. Climatic control is, perhaps, the most likely explanation, because the graded couplets can then be attributed to the annual climatic cycle , and the micrograded laminations to minor fluctuations in sedimentation.

Pettijohn (1936, p. 623) describes "varve-like couplets" in the Abram series of northwestern Ontario as follows: "The lower half of each (couplet) is a white-weathering siltstone of fine grain, which grades upward to a dark, very fine-grained slate. The two halves of each couplet are approximately of the same thickness, though in the thicker layers the coarse silt portion, often sandy in the thickest layers, is predominant." Pettijohn notes that the silty layers are commonly finely laminated, and that the average thickness for the couplets is two to three centimeters. He discusses (p. 624) the possibility of seasonal origin, and concludes: "the sedimentation was cyclical, and, as the resemblance to known annual deposits of Pleistocene age is so striking, a seasonal character is most probable."

Simonen and Kouvo (1951) report abundant "varved schists" in the Bothnian rocks of southwestern Finland. Some varves are as much as

2 meters thick, but in about one-half the exposures, the couplets average less than 0.8 cm thick. For the thin-bedded "varved schists" they report (p. 96): "The fine-grained, light colored siltstone grades upward into the dark-colored, very fine-grained slate, and the demarkation between the siltstone and the slate (at the top of the slate) is usually sharp". Simonen and Kouvo note the similarity of the Finnish graded beds and Pleistocene glacial varves, and they conclude that the couplets in the Bothnian rocks are annual deposits.

Belyea and Scott (1935, p. 225-239) describe "rhythmical banding" in the 1500-foot thick Halifax formation of Nova Scotia, as an alternation of gray quartzite beds and darker slates. They state (p. 228): "The typical quartzite beds are composed of small irregularly shaped crystals of quartz, which range in diameter from an average of 0.04 mm in some beds to 0.10 mm in other beds. These crystals are held together by a fine sericitic cement which seemsto be of the same composition as the groundmass of the slaty member". They note that the quartzite layers are commonly thicker than the slate layers, and that small-scale cross-bedding is abundant in the quartzites. They concluded (p. 238) that the "sediments were laid down in a broad shallow geosyncline. The material was brought down by rivers and deposited at a considerable distance from shore. The banding was caused by the seasonal variation in the size of grains of detritus deposited ... it was acted on by waves and currents, developing a current ripple type of cross-bedding. The uniformity of condition was maintained by a gradual sinking of the sea floor".

Collins (1925, p. 42) describes the "argillite facies" at the base of the Mississagi quartzite as follows: "The argillite is regularly laminated in layers from 1/16 to 1/2-inch thick. The upper and lower

parts of each layer are slightly different in color, giving the formation a finely laminated appearance.

"Under the microscope it appears as an extremely fine-grained aggregate of colorless mica shreds and quartz particles ... Its banding appears to be due not only to the very fine stratification but also to the fact that each lamina becomes progressively finer grained and more argillitic in composition from bottom to top. The lower three-fourths or so of each lamina carries a perceptible percentage of quartz particles embedded in the paste of secondary micas, etc., and the upper part is apparently wholly of this argillitic paste. ... In some laminae the change from bottom to top is gradual; in others, abrupt enough to give the appearance of two distinct parts. Occasional quartz grains 0.05 mm across are scattered here and there through the lower part of the coarser laminae, but as a rule they are not over 0.01 mm. The argillitic materials are still finer grained".

The rocks described above closely resemble the Marion Lake siltstones in showing rhythmic graded bedding of approximately the same scale, and they exhibit similar mineralogy and grain size. The striking resemblance to glacial varves has been noted by the students of these rocks, and they all have postulated seasonal variation in climate as the control for deposition of such rocks. Seasonal control is also here advocated for the graded bedding in the Marion Lake siltstones. The hypothesis is supported by the occurrence within the graded beds of micrograded laminations, which probably reflect minor fluctuations in sedimentation during the major annual climatic cycle.

Sauramo (1923), in a study of Pleistocene varves in Finland, found clay and silt fractions to be intimately mixed (symmict) in marine

varves, and separated (diatactic) in fresh-water varves. He attributed symmictis to the electrolytic effect of sea water. He remarked, however, (p. 94) that coagulation can be retarded or prevented by protective colloids, that the rate of flocculation is a function of the member of particles per unit volume, and that the nature of the electrolytes may greatly affect the process.

The symmictic texture does not imply a lack of grain gradation. Sauramo states (p. 82): "The symmixis is of a variable degree. The band of coarser material ... may be thicker or thinner and constitute a larger or smaller part of the whole. The arrangement of grain in that part is the same as in a diatactic varve".

Silty layers of the Marion Lake banded siltstones contain as much as 45 percent micaceous minerals, representing recrystallized clay minerals. The layers are therefore symmict, and a marine environment of deposition is suggested.

Keco Formation

The name, Keco formation, is here proposed for the sequence of metavolcanic rocks situated stratigraphically above the Marion Lake formation (Table 1). The type locality is on the southwest shore of Keco Lake.

Description

The Keco comprises numerous, very fine-grained, pale greenish gray, rusty- to gray-weathering flows. Pillows, 1 to 5 feet long and rimmed by chloritic selvages 0.5 to 1 inch thick, occur in about 20 percent of the Keco outcrops. Because many of the pillows are balloon-

shaped (Shrock, 1948, p. 364), the upper sides of flows can commonly be determined. Glacially polished surfaces show excellent pillow cross-sections, but in most outcrops, cross-sections suitable for top determinations are found only by extensive stripping of moss cover. All attitude determinations indicate conformability with the overlying and underlying sedimentary rocks.

The upper portions of the steeply-dipping flows weather deeply, forming valleys which appear as prominent lineaments on aerial photographs. Parallelism of the lineaments indicates that the flows are fairly uniform in thickness. Some flows are possibly as much as 600 feet thick, but most of them are less than 100 feet thick.

Thin sections of Keco rocks show abundant acicular feldspar laths, as much as 1 mm long, but less than 0.1 mm thick, randomly oriented in a very fine-grained felted groundmass. Because such delicate feldspars would be best preserved in an originally glassy groundmass, the primary texture was probably intersertal or hyalopilitic.

Composition and Origin

Albite and clinozoisite replace original calcic feldspar. Chlorite, actinolite, and clinozoisite form the fine-grained groundmass, and probably replace volcanic glass. Small amounts of sphene, magnetite, ilmenite, leucoxene, carbonate, quartz and sulphide minerals are invariably present.

Accurate modal analyses cannot be made because the matrix minerals are very fine-grained and intimately intergrown. In Table 9, the chemical composition of a representative Keco sample is compared with the chemical compositions of typical basalts and andesites (Nockolds, 1954,

TABLE 9

Chemical Composition of a Keco Metabasalt Sample
Compared with Average Chemical Compositions of Volcanic Rock Types

	1	2	3	4	5	6	7	8
SiO_2	49.6	50.83	47.90	45.78	43.69	51.33	51.43	47.63
TiO_2	1.0	2.03	1.65	2.63	2.12	1.10	2.60	2.84
Al_2O_3	15.7	14.07	11.84	14.64	9.06	18.04	13.05	14.57
Fe_2O_3	0.5	2.88	2.32	3.16	3.46	3.40	3.36	3.97
FeO	10.19	9.06	9.80	8.73	9.43	5.70	9.74	7.83
MnO	0.2	0.18	0.15	0.20	0.16	0.16	0.19	0.18
MgO	7.4	6.34	14.07	9.39	19.68	6.01	5.28	7.25
CaO	8.7	10.42	9.29	10.74	9.18	10.07	8.78	9.48
Na_2O	3.6	2.23	1.66	2.63	1.49	2.76	3.18	3.75
K_2O	0.1	0.82	0.54	0.95	0.69	0.82	1.04	1.20
H_2O	3.18	0.91	0.59	0.76	0.74	0.45	0.87	0.78
P_2O_5	0.0	0.23	0.19	0.39	0.30	0.16	0.48	0.52
CO_2	0.16							

1. Keco metavolcanic sample, 57-27A. (Analyst: G. Bender, Geol. Surv. Canada)
2. Normal tholeiite (137).
3. Tholeiitic olivine basalt (28).
4. Alkali basalt (96).
5. Alkali-olivine basalt (31).
6. "Central basalt (56).
7. Tholeiitic (basaltic) andesite (26).
8. Alkali andesite (37).

Average compositions, based on number of analyses shown in parentheses; Nockolds, 1954

p. 1019-1020). The Keco sample has low potash and titania contents, as do most of the basalt types, and the silica content is similar to that of the normal tholeiite. However, in lime and soda contents, the sample closely approximates the tholeiitic (basaltic) andesite. The overall chemical composition suggests that the original lava was intermediate in composition between a tholeiitic basalt and basaltic andesite. Chemical composition may have changed somewhat as a result of metasomatism during metamorphism, but because of the close approach to basaltic composition, the most important change was probably addition of water.

Denault Formation

The well-exposed Denault formation consists almost entirely of dolomite and siliceous dolomite. A thin layer of black slate, which crops out at two places on the east shore of Keco Lake, possibly is the basal member of the formation.

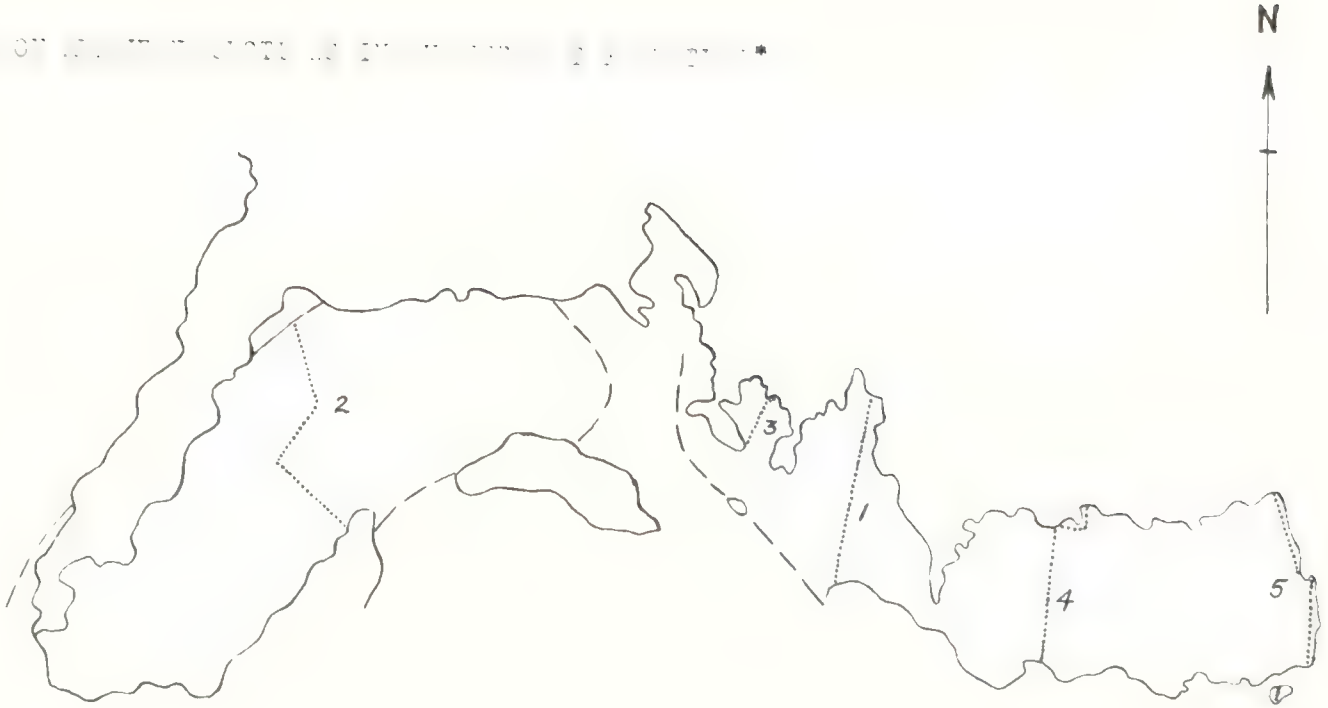
Most of the rocks are very fine grained and would be classified, according to the system of Folk (1959), as dolomicrites. Primary sedimentary structures are abundant and amazingly well-preserved. They can best be seen on weathered surfaces or polished surfaces which have been etched with dilute hydrochloric acid. The specimens etch well, not because of the presence of calcite, but because of grain-size differences related to the structures.

Distribution of lithologies

The following rock types are recognized in the Denault formation: finely laminated siliceous dolomite, massive dolomite, intraformational conglomerate, and stromatolitic dolomite. To determine the

Fig. 6

Locations of the geologic sections measured in the Denault formation.



*For location within the map-area, refer to geologic map, pocket at back

1/2 0 1

SCALE OF MILES

distribution and relative abundance of these rock types, five sections were measured and studied in detail (Fig. 6). On the basis of the detailed work, the Denault was differentiated into 19 lithologic units, most of which are laterally persistent within the map-area. Section descriptions, accompanied by the measured thickness and percentage exposure for each unit, are tabulated in Appendix C, and the data are summarized in a stratigraphic cross-section (pocket at back). The top of Unit 17 is used as a horizontal datum, because this unit can be readily recognized in all of the measured sections.

The measured section data indicate the following distribution of lithologies in the Denault:

Lithology	Percent*
Finely laminated dolomite	30
Massive dolomite	20
Stromatolitic dolomite	45
Intraformational conglomerate	5
Total	100

*Average for exposed rock in the 5 measured sections.

Finely Laminated Siliceous Dolomite

Description

Finely laminated siliceous dolomites, characterized by tabular beds 0.5 to 10 mm thick, compose approximately 30 percent of the Denault formation. Fresh surfaces are cream to light gray, and weathered surfaces orange to yellowish brown. The siliceous dolomites are even textured,

finely crystalline, and break with a smooth conchoidal fracture. Tabular laminations are readily apparent in outcrops, because selective weathering has produced prominent ribbed surfaces (Plate 8A). Thin sections and etched, honed surfaces show that many of the laminations are graded, and that some of the thicker, silty beds are cross-stratified (Plate 8B). The cross-stratified units and basal portions of the graded laminations contain as much as 50 percent quartz grains 0.5 to 0.1 mm in diameter. The grain size and quartz content decrease gradually towards the tops of the graded laminations; in the upper one-third, the grains are less than 0.01 mm in diameter and the quartz content is less than 15 percent.

The quartz grains are rounded to subangular, and some exhibit detrital outlines within overgrowth rims. Silt-sized grains of plagioclase, biotite, and muscovite can be seen in most laminated dolomite thin sections. Zircon, tourmaline, pyrite, and leucoxene are common accessory minerals which occur as detrital grains less than 0.01 mm in diameter. Traces of fine yellowish-brown iron oxides, probably derived from the dolomite as a result of weathering, are dispersed through most specimens. The dolomite is pale yellow to colorless, mostly untwinned, and occurs in subhedral, subspherical grains identical in size to the terrigenous silt with which it is associated. Grain gradation of the dolomite corresponds to grain gradation of the clearly detrital silicates, and is of the normal, waning current type (Pettijohn, 1957, p. 171).

Origin

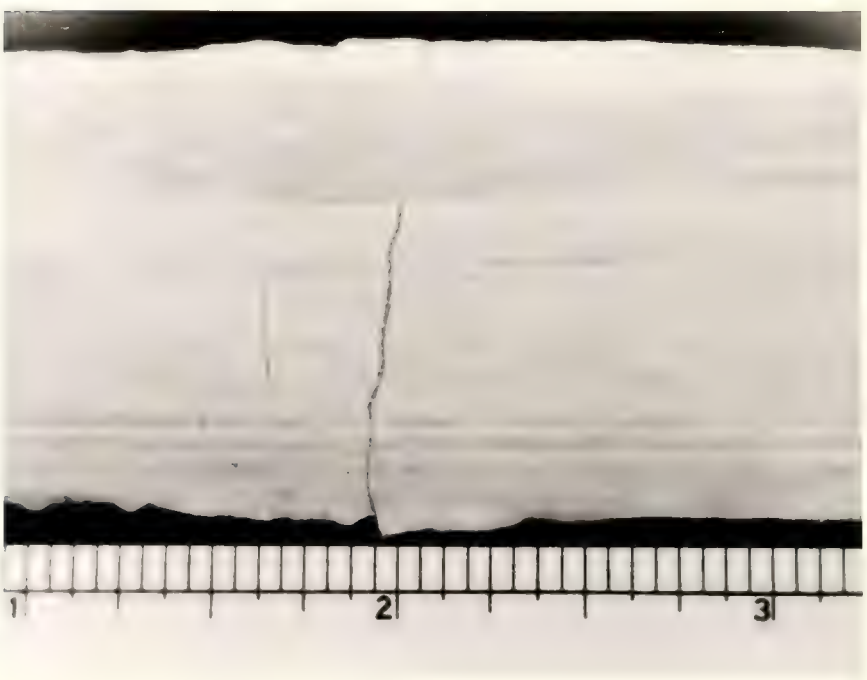
Cross-bedding and graded bedding indicate detrital origin for the finely laminated Denault beds. The fine grain size and preservation of lamination suggest deposition in relatively quiet water. Erosion of

PLATE 8

Finely laminated siliceous Denault dolomite

A. Outcrop of laminated siliceous dolomite showing ribbed weathered surface.

B. Polished surface of laminated siliceous dolomite showing graded beds (bottom) and cross-laminated unit (top).


1
2
3

pre-existing formations may have yielded the fine detritus of the laminated Denault beds, but no older dolomites occur in either the Labrador Trough or the basement rocks. Direct precipitation within the basin of sedimentation is therefore the likely source of the fine-grained dolomite. Erosion and comminution of older Denault beds may have contributed some sediment to Unit 19, but such a process cannot be invoked for the lowermost beds of Unit 1, which also are finely laminated. Probably much of the dolomite in the finely laminated Denault beds represents precipitated muds which, because of some transport prior to deposition, have formed beds showing graded-bedding and cross-bedding.

Massive Dolomite

Description

Fine- to medium-grained massive dolomite composes approximately 20 percent of the Denault formation. It is tough, light gray, dark-weathering, and breaks with an irregular fracture. Joints are prominent secondary structures, and indistinct, widely-spaced stratification can be seen in some outcrops (Plate 9A). Beds of intraformational conglomerate occur in several massive units. Veinlets, lenticular pods and irregular layers of quartz and chalcedony are abundant.

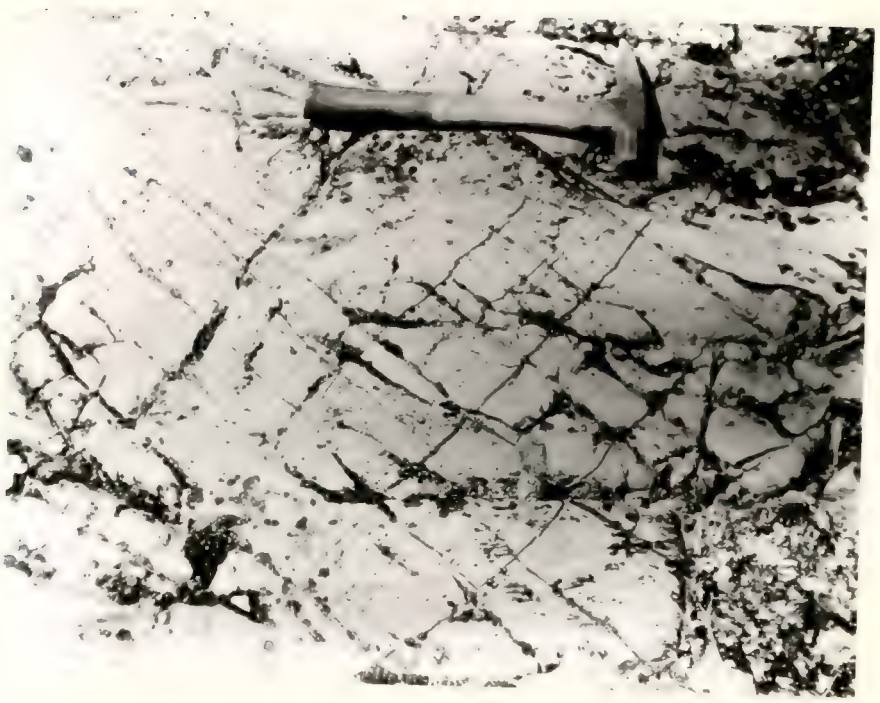
The massive dolomite contains less than 10 percent quartz and only traces of other silicate minerals. Most of the quartz occurs as scattered anhedral grains less than 0.5 mm in diameter, but quartz grains as much as 1 mm in diameter are dispersed through the upper part of Unit 8. The larger quartz grains commonly show well-rounded detrital outlines. Most of the dolomite occurs as colorless anhedral grains, 0.2 to 1.5 mm in diameter. Some grains show rounded outlines suggestive of clastic

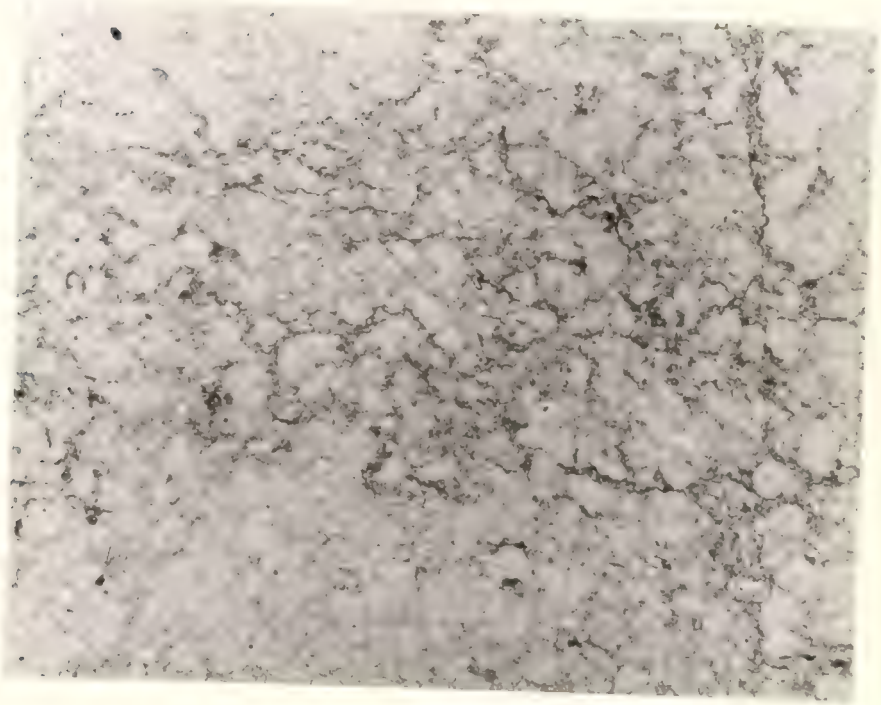
PLATE 9

Massive Denault dolomite

A. Outcrop showing weathered surface of massive dolomite. Note the two prominent joint sets and lack of offset along the joints. Transverse section, hammer handle parallels bedding.

B. Photomicrograph of massive dolomite showing granoblastic texture. Note abundance of microstylolites. Ordinary light, X20.





origin, but commonly they form a granoblastic mosaic in which the original nature of grain boundaries has been obscured by solution along microstylolitic seams (Plate 9B). In some massive beds, sand-sized pellets composed of very fine-grained dolomite are abundant. The pellets are ellipsoidal to subspherical, well rounded, and well sorted. Grain size of the dolomite cement is less than 0.2 mm and that of most pellets is less than 0.005 mm. Some pellets contain dolomite rhombohedra, as large as 0.5 mm in diameter, which have apparently grown at the expense of the microcrystalline dolomite. All stages between pelletoid and granoblastic dolomite can be seen in thin section, indicating that the granoblastic rocks are recrystallized dolomite sands. The control of recrystallization is not known, but abundance of microstylolites in the granoblastic dolomites shows that pressure and access of solutions probably were important factors.

Origin

The massive dolomites are clastic, partly recrystallized, dolomite sands. Good sorting of the detrital particles suggests deposition in a turbulent, shallow water environment. Penecontemporaneous reworking of cohesive dolomite muds, perhaps hardened by subaerial exposure, may have yielded the dolomite sands. At least some reworking is suggested by the occurrence of intraformational conglomerates in the massive units. The dolomite sand grains could also have formed by aggregation and cementation of fine-grained muds. Such origin has been demonstrated for modern carbonate sands in the Bahamas (Illing, 1954, p. 26).

Stromatolitic Dolomite

Definition and Classification

The term "stromatolite" denotes a laminated structure which is believed to record algal activity (Kalkowsky, 1908, p. 68). Stromatolites are abundant in the Denault formation, and several distinctive types can be recognized. "Stromatolitic dolomite" here refers to lithologies in which stromatolites are dominant structures. Problems of nomenclature did not arise during measurement of the Denault sections, because lithologies containing only a few scattered stromatolites are absent.

Generic and specific names have been applied to stromatolites by many authors. J. H. Johnson, (1946, p. 1089), justifies binomial classification on the basis that: "Different assemblages (of algae) formed colonies sufficiently different to be separated by megascopic characteristics". However, recent studies of algal structures in Florida (Ginsburg, written communication), indicate that different forms can be built by the same species or assemblages; environment is the important control of form in many cases.

There are at least nine stromatolite types which differ in gross form and to which generic names have been applied. Six of them are figured by Cloud, (1942, p. 367), who states, (p. 366), "It seems better to describe the peculiarities of stromatolites in the same manner that one would describe types of chert or textural characters of sedimentary rocks. As a matter of convenience, it is perhaps well to retain such familiar form designations as cryptozoon, gymnosolen, and collenia. These ... should be used in a vernacular sense, without being italicized or capitalized".

Rezak (1957, p. 130), in discussing stromatolites of the Belt Series, states: "I have used the names of form-genera and form-species with the understanding that the classification is purely artifical and is used as a matter of convenience. However, I am aware of the possibility of confusing these names with the names of biologic entities. Perhaps a new method of classifying these structures should be devised ... the new names should not have resemblance to biologic names". The writer agrees with Rezak's latter statement, but questions whether it is advisable to continue the use of biologic nomenclature, while awaiting, as Rezak suggests, (p. 130), a classification "based on ... a comprehensive examination of all kinds of 'algaloid' structures from all parts of the geologic column". Accordingly, tentative non-biologic names are used here to denote the six basic types of stromatolites in the Bonault formation (Fig. 7). Generic names assigned to similar structures are parenthesized in the descriptions which follow:

Hemispherical stromatolites (Collenia Walcott): consist of successive hemispherical, convex-upward laminations. Laminations are of uniform thickness and are roughly concentric with respect to the center of the basal surface.

Bulbous stromatolites (Cryptozoon Hall): these forms are attached at a point, and consist of successive balloon-shaped laminations which are thickest at the top and pinch out near the base.

Columnar stromatolites (Archaeozoon Mathew): cylindrical columns consist of successive hemispherical, convex-upward laminations. Adjacent columns are tightly packed.

Digitate stromatolites (Gymnosolen Steinmann): consist of discrete cylindrical columns which commonly branch upwards into parallel columns of lesser diameter.

Pisolitic stromatolites (Pycnostroma Gurich): individual, subspherical, concentrically laminated bodies.

Undulatory stromatolites (Weedia Walcott): consist of laterally continuous laminations which have irregular wavy boundaries.

Fig. 7

The six basic types of stromatolites in the Denault formation.

A. Hemispherical stromatolites (Collenia).

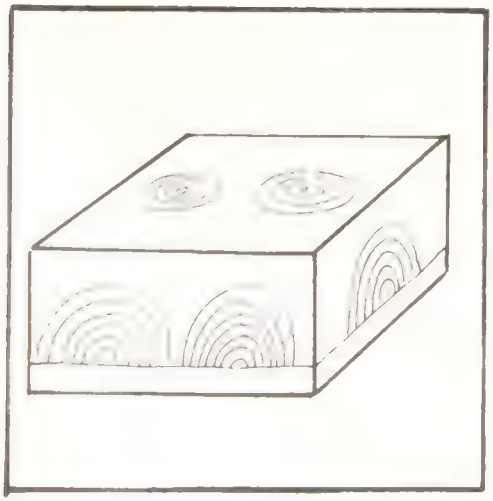
B. Bulbous stromatolites (Cryptozoon).

C. Columnar stromatolites (Archaeozoon).

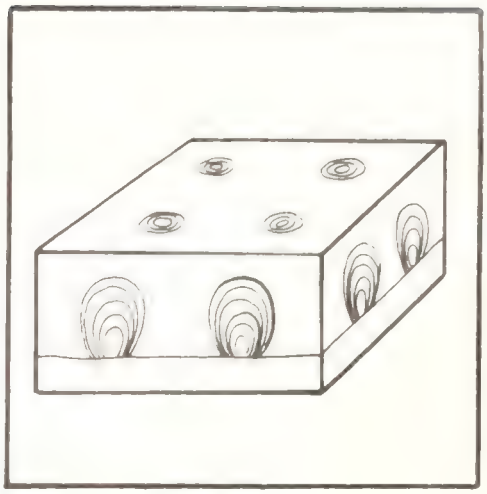
D. Undulatory stromatolites (Weedia).

E. Digitate stromatolites (Gymnosolen).

F. Pisolitic stromatolites (Pycnostroma).



A



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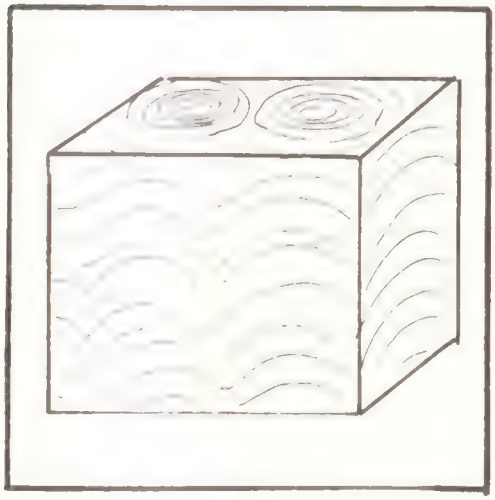

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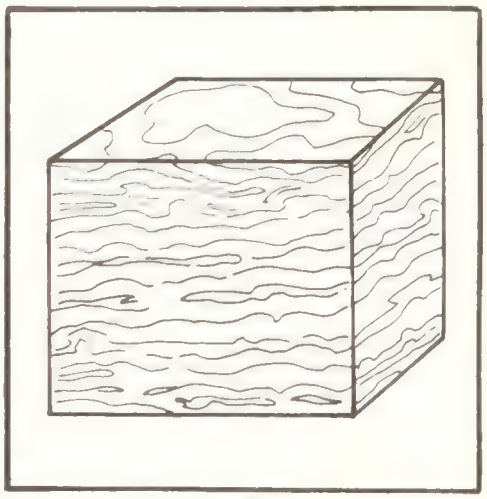

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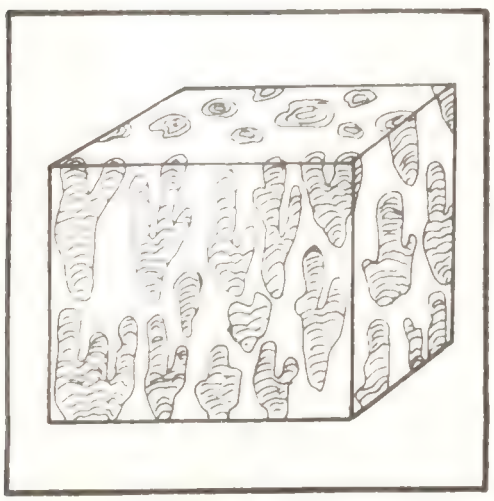

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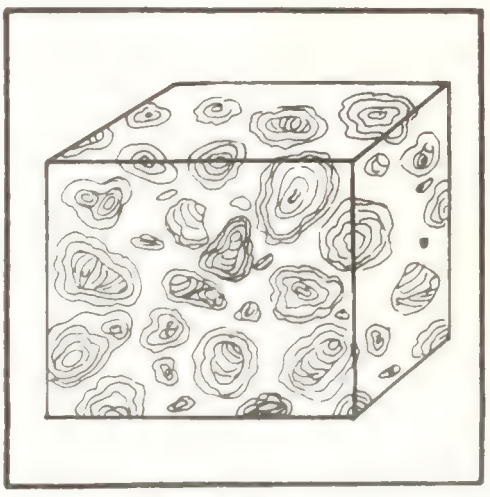


F

Description of the Denault Stromatolites

The measured section data (Appendix C) indicate that more than 45 percent of the Denault formation consists of stromatolitic dolomite. The distribution of the stromatolite types is as follows:

Type	Percent*
Columnar, hemispherical, and bulbous stromatolites	10
Digitate stromatolites	5
Undulatory stromatolites	25
Tiscolitic stromatolites	5
Total	45

*Average for five measured sections.

Columnar stromatolites (Plate 10) occur in five distinct zones in the Denault formation. Diameters of the columns range from 5 to 12 inches in three of the zones, and from 2 to 5 inches in the other two zones (see cross-section, pocket at back). Axes of the columns are perpendicular to the bedding, and adjacent columns are attached to the same bedding surface. Hemispherical and bulbous stromatolites commonly occur with the columnar stromatolites, and are believed to merely represent early growth stages of the columnar type. Indeed, the basal parts of most columnar stromatolites exhibit a lamination intermediate in character between the hemispherical and bulbous types.

The stromatolite laminations, which range from 1 to 20 mm in thickness, are commonly continuous through several adjacent columns. The laminated structure is accentuated by alternation of dark gray to black layers and light gray to buff layers (Plate 12). Thin sections

PLATE 10

Columnar stromatolites

A. Weathered joint surfaces showing longitudinal sections of columnar stromatolites. Scale extended one foot.

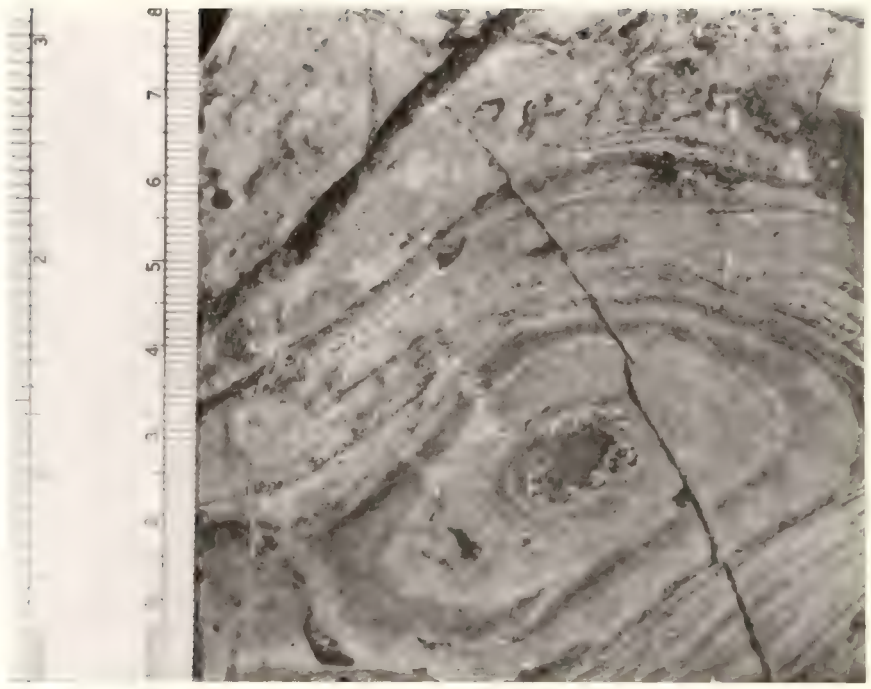
B. Block of stromatolitic dolomite containing two adjacent stromatolite columns. Note alternation of dark and light hemispherical laminations. Scale in lower right corner is two inches long.

PLATE 11

Hemispherical stromatolites

A. Weathered transverse section of hemispherical stromatolite. Darker layers are carbonaceous.

B. Etched longitudinal section of a hemispherical stromatolite. Note fragmented carbonaceous lamination in central part.



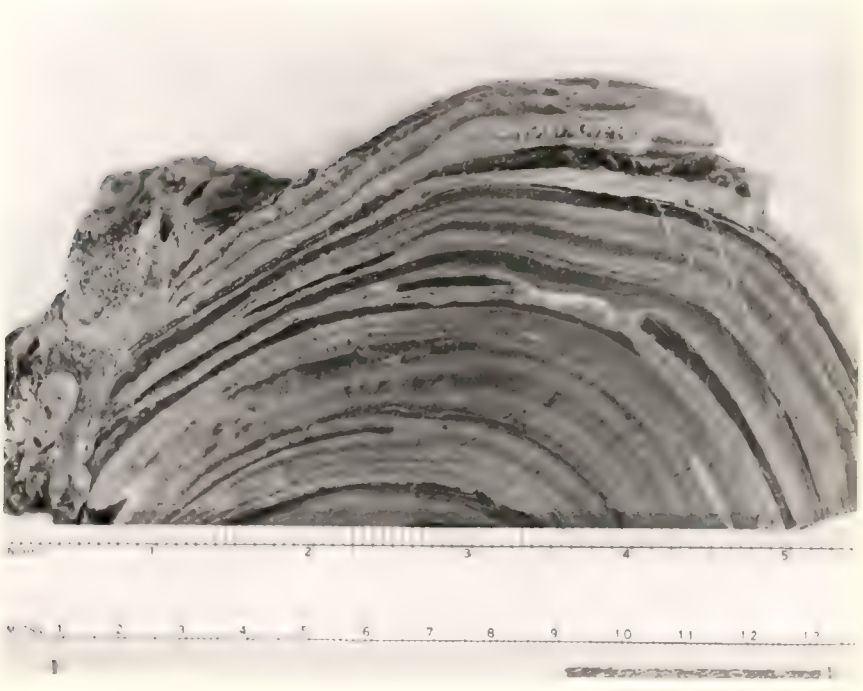


PLATE 12

Photomicrograph of hemispherical stromatolite

Note alternation of dark, carbonaceous layers and less carbonaceous layers which contain abundant quartz silt and structureless pellets. Truncation of the central layer suggests vigorous wave action during growth. Ordinary light, X6.

show that the dark layers are composed of anhedral grains of dolomite 0.001 to 0.01 mm in diameter. Coloration is due to less than 1 percent amorphous carbon which is dispersed through the layers as thin films between dolomite grain boundaries. Grain size in the light-colored layers ranges from 0.01 to 0.05 mm. Fragmental detritus and quartz silt are commonly abundant in the carbon-poor layers, but scarce in the carbonaceous layers. Some carbonaceous layers are sharply truncated (Plates 11B, 12) and the interspaces between columns commonly contain platy carbonaceous fragments which correspond in thickness, texture, and structure to the dark stromatolite layers.

Digitate stromatolites of the Denault form beds less than 1 foot thick. In longitudinal section, these stromatolites appear as finger-like columns 1 to 5 mm in diameter and less than 2 cm in height (Plate 13A). They exhibit domical, convex-upward laminations, 0.05 to 1.0 mm thick, which are marked by thin, crenulate films of carbon. In transverse section the carbonaceous films within individual columns show as somewhat irregular but concentric rings. The columns are separated by interspaces occupied by fine-grained carbonaceous dolomite, which, in places, exhibits a faint concave-upward lamination (Plate 13B). Fragmental detritus and quartz silt are abundant in the interspaces but scarce in the columns. Grain size of dolomite in the stromatolites ranges from 0.001 to 0.05 mm and, in the interspaces, from 0.01 to 0.2 mm. Successive laminations in adjacent stromatolites show a definite correspondence in thickness (Plate 14). The stromatolites are locally closely packed, but in most places they compose less than half of the bed in which they occur.

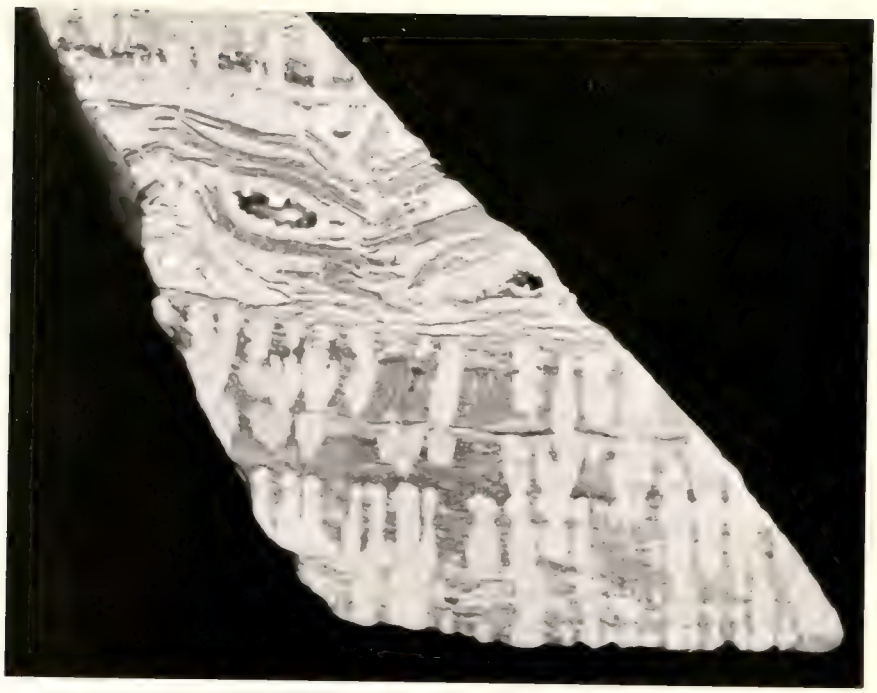
Pisolitic stromatolites, 5 to 15 mm in diameter, form 1- to 5-foot thick beds in the Denault. The pisolitic zones are dark gray to

PLATE 13

Digitate stromatolites

A. Etched surface showing stromatolitic unit overlain by intraformational conglomerate. The fragments of the conglomerate are carbonaceous, and probably represent algae-reinforced layers derived from hemispherical stromatolites (See Plate 12). Longitudinal section (XI).

B. Detail of the digitate stromatolites shown in A. Note upward branching and convex-upward laminations of the "fingers", and faint concave-upward lamination of the detrital matrix. Longitudinal section, scale in inches.

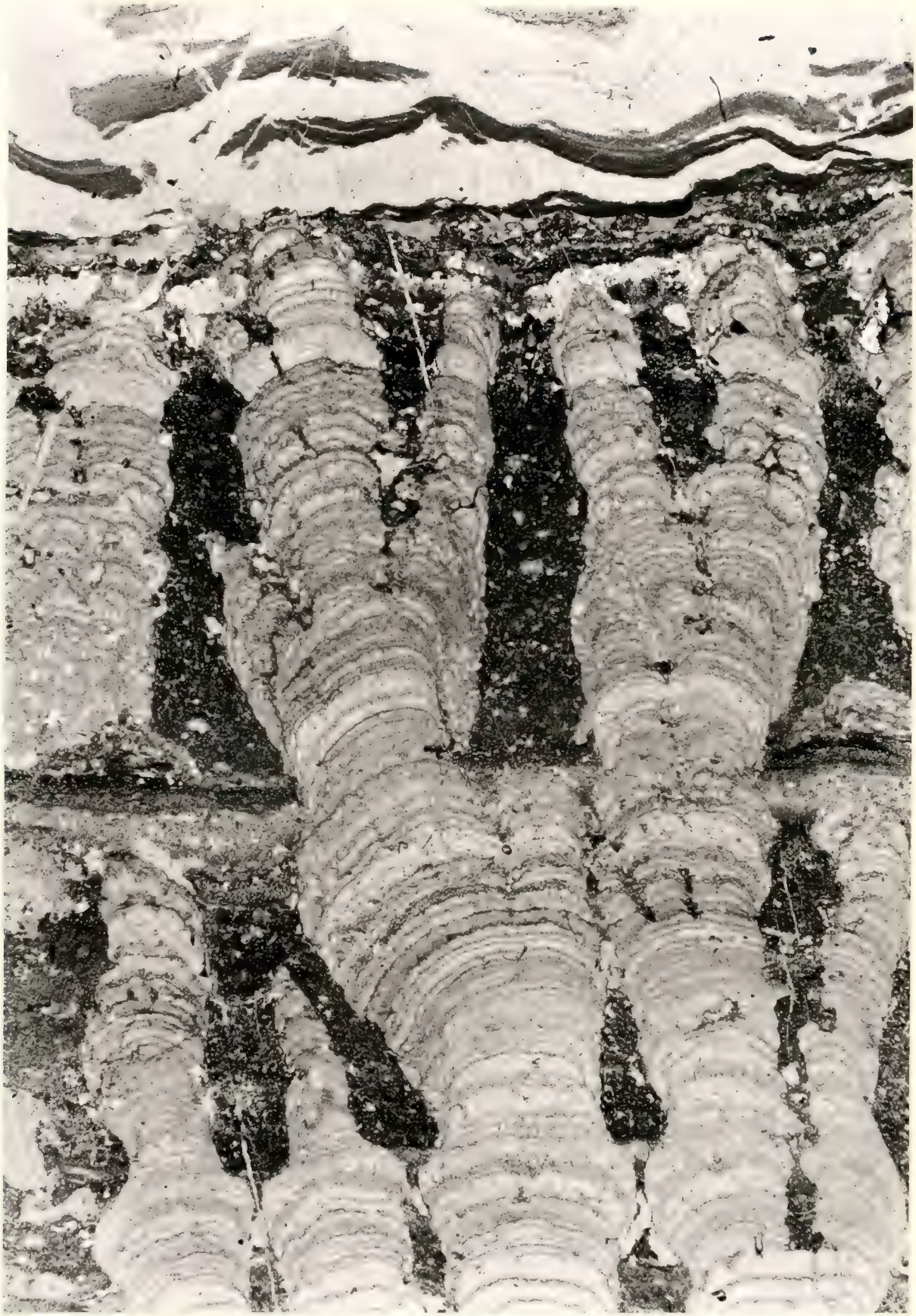


1
2

PLATE 14

Photomicrograph of digitate stromatolites

Thin section showing detrital nature of the interspaces between digitate stromatolites, and the correspondence in thickness for successive laminations in adjacent "fingers". Ordinary light, X8.



black due to a carbon content of as much as 1 percent. The pisolites are composed of anhedral dolomite grains 0.001 to 0.02 mm in diameter, laminated by concentric films of carbon at intervals of 0.02 to 1.0 mm. Structureless pellets, as large as 10 mm in diameter, are invariably associated with the pisolites and some of the pisolites contain a small pellet or quartz grain nucleus. The enveloping laminations are somewhat crenulate, and successive laminations are commonly eccentric. Outer sets of laminations truncate inner sets in many of the stromatolites. Few of the pisolites are spherical; ellipsoidal to irregular "flower-like" forms predominate (Plate 15). In some beds, carbonaceous pellets are more abundant than the pisolitic stromatolites, and fragmental pisolites are common (Plate 16). In some units, digitate and pisolitic stromatolites occur together (Plate 17).

Undulatory stromatolites are very abundant in the Denault formation. They are recognized in the field by the characteristic crinkly-ribbed weathered surfaces (Plate 18A). Little can be seen on fresh surfaces, but careful examination of the weathered surfaces reveals the intricate structures which compose this type of stromatolitic dolomite, and polished surfaces show even more detail (Plate 18B). The continuous undulatory layers occur 5 to 20 mm apart, and are 1 to 10 mm thick. Some are structureless, but most exhibit crenulate lamination. The intervening parts are commonly occupied by pisolites, pellets, fragments, and small columnar stromatolites.

Origin of Stromatolites

Stromatolites are generally recognized to be related to algal activity, but because their organic origin has at times been questioned,

PLATE 15

Pisolitic stromatolites

A. Polished surface of pisolitic stromatolites, showing the diversity of laminated and structureless forms. (X2).

B. Weathered surface of pisolitic stromatolites. Note abundance of structureless pellets between the laminated bodies. (X5).

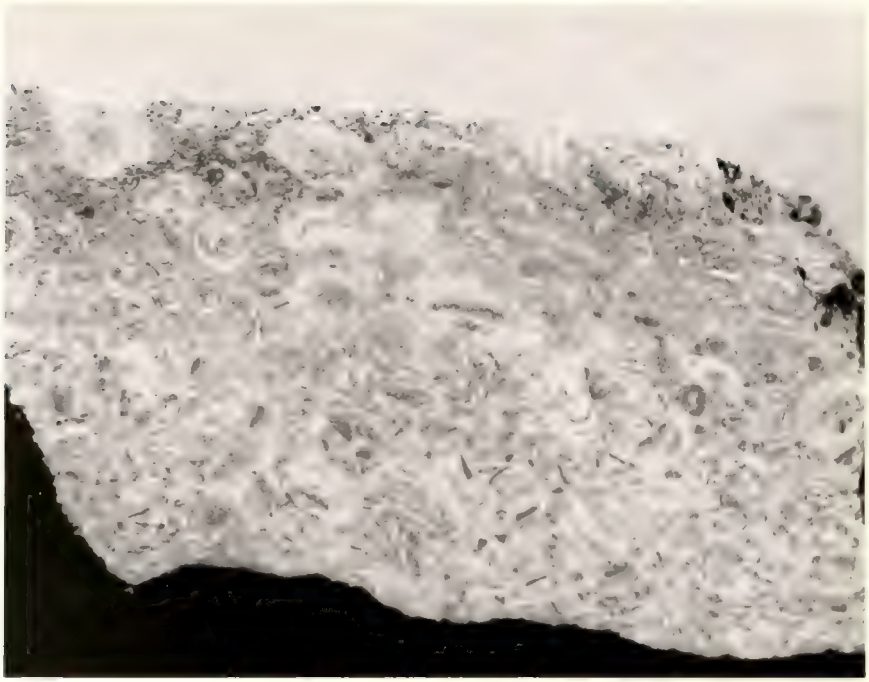


PLATE 16

Photomicrograph of pisolitic stromatolites

Concentrically laminated pisolitic stromatolites and structureless carbonaceous pellets cemented by clear, crystalline dolomite. Note fragmental pisolite near center of photograph. (x8)

PLATE 17

Photomicrograph of pisolitic stromatolites

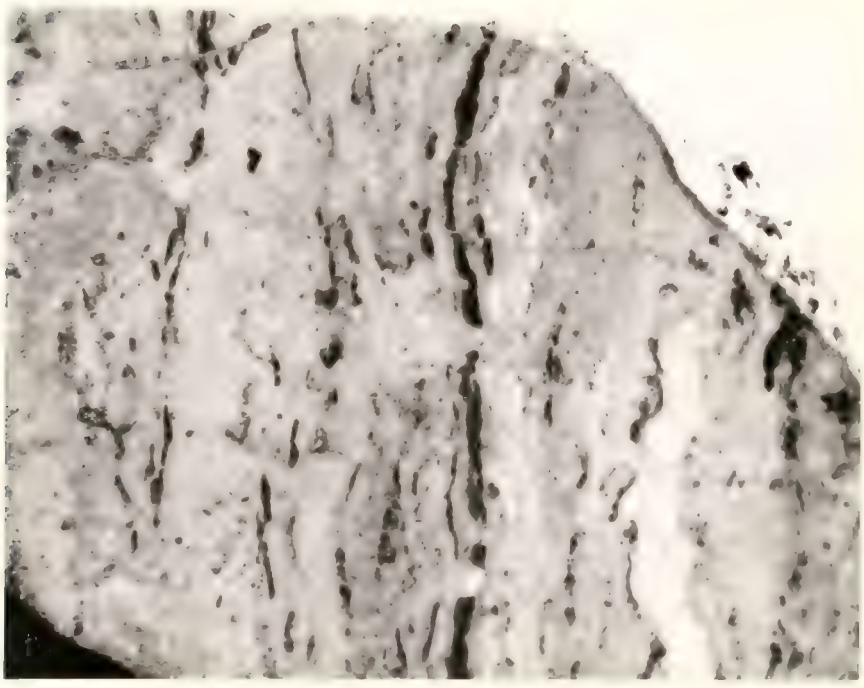
Thin section showing association of pisolitic stromatolites (top), digitate stromatolites (bottom), and fragmental detritus. Note composite nucleus in pisolitic stromatolite in upper right corner. Longitudinal section, ordinary light, X10.

PLATE 18

Undulatory stromatolites

A. Differentially weathered outcrop showing undulatory stromatolites. Note irregular pods of milky quartz near bottom and top of photograph.

B. Polished surface showing discontinuous nature of undulatory stromatolite laminations. Dolomite gray, carbonaceous dolomite dark gray to black, quartz and chalcedony white, X1.5.



possible inorganic origins will be reviewed. Inorganic processes which have been proposed include:

1. Mechanical deformation of consolidated rock.
2. Soft-sediment deformation.
3. Concretionary growth.
4. Inorganic precipitation.

1. Young (1943, p. 94) describes rocks of the Dolomite Series of South Africa which contain "domical-columnar structure" as: "... an assemblage of rocks whose members have adjusted themselves to compressive stresses in diverse manner according to their individual characters and circumstances, and that the result has been a shortening of the section in every direction." He also states (p. 92): "No attempt will be made to analyse the mechanics of any of the deformation described." His argument for mechanical origin is based mainly on the presence of breccia zones between adjacent columns. However, he does not offer evidence to show that the "breccias" are truly of secondary origin. Further, in two figures shown by Young, lamination is clearly continuous between the columns (Young, 1943, Plate 22, Fig. 2; Plate 24, Fig. 2). Such continuity of lamination is difficult to reconcile with "mylonitization between the columns" (Young, 1943, p. 98). The possibility of mechanical doming should be considered for structures of the columnar stromatolite type, but it cannot be invoked for the digitate and pisolitic forms.

2. Schneiderhöhn (1921, p. 264) suggested that submarine gliding of thin muddy layers could result in forms similar to algal structures. Fairbridge (1946) described several structures which may form by submarine slump of soft, unconsolidated sediments. The type of deformation described

as "sealing-wax flow" (Fairbridge, 1946, p. 87) could perhaps give rise to forms similar to stromatolites of the undulatory type. Soft-sediment adjustment between water-charged beds under the influence of sedimentary loading, and differential compaction as a result of compositional inhomogeneities are other possible modes of origin for irregular lamination.

3. Pisolites of bauxite deposits are commonly assigned to concretionary origin (Gordon and Tracey, 1952, p. 27), and pisolites in limestones have also been interpreted as concretions by some workers (Swineford et al, 1958, p. 115). Because true concretions (Pettijohn, 1957, p. 197) are the product of in situ, post-depositional accumulation of mineral matter around a nucleus, successive layers form concentric shells which transect bedding. Algal pisolites, however, grow prior to deposition in beds, and therefore cannot transect bedding. Further, because algal pisolites form at the depositional interface, they are subject to water-wear and fracture. Fragmental pisolites might therefore be expected in pisolitic deposits of algal origin. However, fragmental pisolites can also be expected in pisolitic deposits which form by inorganic accretion at the depositional interface.

4. The tufa domes of Lake Lahonton, which in gross form resemble the hemispherical types of stromatolites, were attributed to chemical precipitation by Russell (1885, p. 188). More recently, algal origin for at least some of the Lahonton tufas has been demonstrated (Jones, 1925, p. 7). Emig noted the common association of algae and tufa (or travertine), but, with reference to the deposits in Yellowstone Park, stated (1917, p. 59): "... there is no question but that travertine would be formed from the hot mineral water even if the plant life were absent. The

vegetation in this particular instance, by its presence, acts only as a passive agent in determining the form and structure of the deposit, and does not necessarily take an active part in the precipitation of calcium carbonate."

A South African occurrence of travertine, apparently formed by inorganic precipitation of hot spring waters, is reported by Kupferburger (1935, p. 10) as showing "irregular, wavy, horizontal stratification." The deposit, which is approximately one half mile in diameter and as much as 200 feet thick, is described as a "crater" which was built up by evaporation of carbonate-rich water released from a central spring.

Ohle and Brown (1954, p. 209-210) described "fingering" in "snurly rock" of the Bonterre dolomite. They at first interpreted the "fingers" as the product of solution (1954, p. 209), but in a later paper (Ohle and Brown, 1954, p. 935-936), they reinterpreted the "fingers" as algal structures.

Caliche, a lime-rich deposit formed in the soils of semiarid regions (Pettijohn, 1957, p. 410), commonly contains structures which resemble pisolitic and undulatory stromatolites. Some structures in caliche have, in fact, been attributed to algal activity (Price, Elias, and Frye, 1946). Bretz and Horberg, however (1949), have convincingly shown that the structures can be explained by soil-forming processes alone. Important features of caliche include physiographic rather than stratigraphic control, and development of the typical caliche profile (Bretz and Horberg, 1949, Fig. 2).

Origin of the Denault Stromatolites

No inorganic processes can adequately account for all of the Denault stromatolites. The obvious evidence for primary origin, including truncation of laminations, intraformational conglomerates, and restriction to stratigraphic zones, eliminates the possibility of post-consolidation deformation. Soft-sediment deformation could perhaps give rise to structures of the undulatory stromatolite type, but could not produce the pisolitic or digitate forms. Because true concretions are secondary structures, the arguments for primary origin suffice to eliminate concretionary origin. Inorganic precipitation alone can perhaps lead to formation of both pisolites and deposits showing irregular lamination, but unless accompanied by organic activity, can hardly result in branched or domical structures. The most compelling argument in favour of organic origin, however, is the carbonaceous nature of the stromatolite zones. Further, modern structures of striking similarity to the Denault stromatolites are known to be forming today as a direct result of algal activity.

The pisolitic, columnar, and digitate stromatolites of the Denault formation are interpreted as algal structures. They occur in stratigraphically continuous zones which probably are biostromes. The possibility of inorganic origin for the undulatory stromatolites is recognized, but intimate association with the other stromatolites strongly suggests algal origin.

Environment and Mode of Growth

Because stromatolites occur in about 45 percent of the Denault, interpretation of their significance is of fundamental importance. Modern

algal analogues have been studied and described in considerable detail; hence comparisons are possible.

Black (1933, p. 169), found that algal structures with characteristic morphology form only above low water mark, although algae serve to bind sediments at greater depths without producing any internal structure. According to Ginsburg (1959, written communication), discrete laminated forms seldom grow at depths greater than 15 feet. Cloud (1942, p. 371) notes that because algae need light for photosynthesis, they seldom can grow profusely at depths greater than 10 meters. For the Denault stromatolites, the association with intraformational conglomerates, the common occurrence of wave-broken and wave-worn laminations, and the abundance of stromatolitic detritus are strong evidence for shallow water conditions.

Modern algae grow in arctic to tropical waters, and hence provide no definite evidence of environmental temperature. However, as pointed out by Cloud (1942, p. 370), algae seem to grow better in warm water than in cold water. The abundance of stromatolites in the Denault is at least suggestive, then, of warm-water deposition.

Modern algal structures are abundant in both shallow-marine and lacustrine environments. They thrive in sea water (Ginsburg, 1958, p. 311), fresh water (Black, 1933, p. 169), and hypersaline water (Bardley, 1938, p. 1395). Cloud (1942, p. 372) at first thought that digitate stromatolites might indicate non-marine origin, but later (1945, p. 108) noted three occurrences of digitate stromatolites in marine strata. Algal structures similar to stromatolites of the pisolitic type grow in modern fresh water environments (Bradley, 1929, p. 219; Clarke, 1900; Murray, 1895; Roddy, 1915), but Twenhofel (1919) and Sando (1957, p. 38)

describe pisolitic stromatolites in ancient sedimentary rocks of known marine origin. Algae which produce forms of the hemispherical and undulatory type are now active in marine, brackish, and fresh waters in the Bahamas (Black, 1933, p. 169). Clearly, then, no specific type of stromatolite can be used as a salinity index, because all types of algal structures can form in both continental and marine environments.

Algae can contribute to carbonate deposition by active precitation of lime, but their greatest contribution appears to be the fixation of sedimentary particles. Black (1933, p. 186) described algal mats in the Bahamas which trap fine detritus swept in by currents. Ginsburg (1958, p. 311) reported similar mats along the Florida coast, and described an experiment in which an algal culture, when covered by a layer of sediment 4 mm thick, grew through the layer and re-established a surface mat within 24 hours. The alternation of carbon-rich and carbon-poor layers in the Denault columnar stromatolites is suggestive of sediment-trap origin, the carbonaceous layers representing alga-rich layers which formed during periods of quiescence, and the carbon-poor layers representing detritus swept in during periods of turbulence. Strong support is lent to this mode of origin by the observation that dolomite fragments and grains of quartz silt are abundant in the carbon-poor layers, but scarce in the carbon-rich layers.

The pisolitic and digitate stromatolites show layering similar to that of the columnar stromatolites, but the layers contain only minor fragmental dolomite and quartz silt. This suggests that growth of the pisolitic and digitate stromatolites occurred mainly by direct precipitation rather than by trapping of sediment. Concentrically laminated bodies similar to the pisolitic stromatolites of the Denault are formed

today by lime-precipitating algae (Murray, 1895; Clarke, 1900; Roddy, 1915), and modern algal structures analogous to the digitate stromatolites have been shown to grow in the same way (Bradley, 1929, p. 215; Pia, 1933, p. 147).

The pisolitic stromatolites grew unattached, as indicated by the complete envelopment of successive laminations around a nucleus. Truncation of laminations and the common occurrence of large pisolite fragments in the matrix demonstrate rough water conditions. The digitate stromatolites, however, exhibit no truncation of lamination, and their parallel, upward-branching habit demonstrates growth from a fixed base. Fragmental detritus commonly occupies interspaces between the digitate stromatolites, but it is much finer grained than detritus associated with the pisolitic stromatolites. The digitate stromatolites evidently developed in quieter waters than did the pisolitic stromatolites. Perhaps, then, both types were formed by the same algae, and the intensity of turbulence controlled the morphology. Lamination in both pisolitic and digitate stromatolites is marked by thin carbonaceous films which exhibit delicate crenulate outline, suggesting similarity of origin.

Undulatory stromatolites are the most abundant, but least distinctive forms in the Denault. Fine detritus is commonly present in the irregular laminations, suggesting that if algae were active, they were probably of the sediment-binding type. The common occurrence of other stromatolites in the undulatory zones indicates at least some algal activity. Algal mat deposits currently forming on tidal flats in Florida and the Bahamas exhibit irregular lamination, and Sando (1957, p. 35) has interpreted portions of the Beekmantown group in Maryland as possible algal mat deposits on the basis of similar lamination.

Intraformational Conglomerate

Description

Beds of intraformational conglomerate, as much as 3 feet thick, compose about 5 percent of the Denault formation. According to Walcott, (1894, p. 192), an intraformational conglomerate is "formed within a geologic formation of material derived from and deposited within that formation", and it is in this sense that the term is here used. Some Denault conglomerates separate individual beds of stromatolites or occur at the tops of the stromatolitic zones; others form beds a few feet thick in the massive units.

The intraformational conglomerates associated with the stromatolites consist of thin, curled, dark carbonaceous dolomite fragments, as much as 3 cm long, in a light-colored matrix of silt-sized dolomite grains (Plate 13A). The fragments are angular, and many have length to width ratios of more than 20:1. They are commonly laminated and, under the microscope, exhibit structure and texture identical to dark layers of the hemispherical stromatolites.

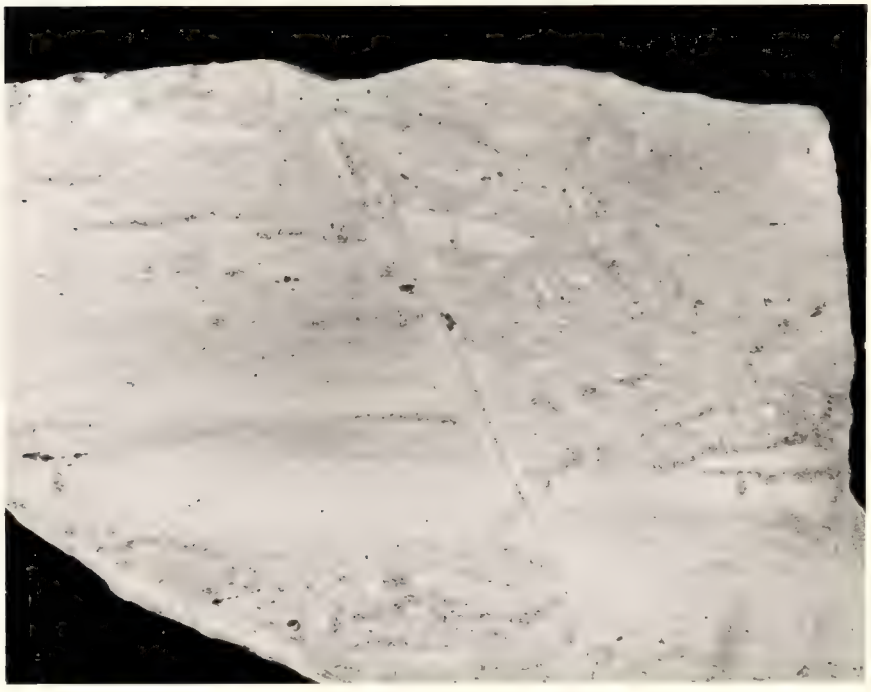
The intraformational conglomerates in the massive dolomites consist of poorly sorted, subangular to rounded dolomite fragments in a matrix of sandy or pelletoid dolomite (Plate 19). The fragments, which are as much as 5 cm long, range from subspherical to markedly tabular. Some of the flat pebbles have length to width ratios of more than 10:1, and, in places where such platy fragments are abundant, imbricate structure can commonly be seen. Beds of the conglomerate are commonly less than 3 feet thick, but are laterally continuous for as much as a few hundred feet. In thin section, many pebbles are seen to be composed of

PLATE 19

Intraformational conglomerate

A. Weathered surface of intraformational conglomerate. (X2).

B. Polished surface of intraformational conglomerate. Note abundance of flat pebbles. (X2).



microcrystalline dolomite, but some are composed of fine-grained pellets identical to those in unrecrystallized portions of the massive beds. Pebbles, sand grains, and pellets are cemented mostly by very fine-grained dolomite, but small patches of dolomite grains 0.1 to 0.5 mm in diameter are common. None of the components are carbonaceous.

Origin

The intraformational conglomerates are clearly of sedimentary rather than tectonic origin, because they are confined by stratigraphic boundaries, the matrix is commonly sandy, and many of the fragments show water-wear. The conglomerates are of local derivation: phenoclasts (Field, 1916, p. 32) in conglomerates associated with the massive dolomite commonly show the same texture as pellets in the massive beds, and phenoclasts associated with the stromatolitic dolomite are carbonaceous and possess lamination identical to that of the stromatolites. That some of the conglomerates show edgewise arrangement indicates strong current or wave action.

Flat pebble intraformational conglomerates are commonly assigned to mud-crack origin. Walcott (1894) first proposed this hypothesis as an outcome of observing desiccation on a modern tidal flat at Noyes Point, Rhode Island. Field (1916, p. 52) summarized the concept as follows: "A tidal flat whose sediments were composed of a limy mud, when desiccated, would, if disturbed by subsequent and sufficiently powerful tidal action, yield a quantity of tough, not brittle, phenoclasts which might be redeposited with little or no signs of attrition except at the edges."

The writer favours the mud crack origin for the Denault conglomerates in the massive dolomite, because it offers a method of separating thin, flat slabs from an otherwise homogeneous deposit. For the conglomerates associated with stromatolites, however, primary structure of the columnar stromatolites may have been sufficient to control fragmentation of thin flat slabs, in which case the desiccation process need not be invoked. The structure, as visuallized by comparison with modern algal forms, consisted of layers bound by algal filaments, alternated with layers relatively free of algal filaments. Subjected to strong wave action, the algae-reinforced layers yielded tough fragments more resistant to attrition than the intercalated layers of sediment. The occurrence of abruptly terminated carbonaceous layers in some of the stromatolites (Plate 12) substantiates such origin.

Composition of the Denault

The chemical and normative compositions of a laminated siliceous dolomite sample and a stromatolitic dolomite sample are shown in Table 10. The laminated siliceous sample is representative of Unit 1, and the stromatolitic sample represents a carbon-rich layer of a columnar stromatolite.

Dolomite is the only carbonate mineral in the two Denault samples, according to the normative calculations, and x-ray diffraction of 20 samples representative of other units failed to detect calcite. Notable is the presence of carbon in the sample of stromatolitic dolomite, and the abundance of quartz and mica in the sample of laminated siliceous dolomite.

TABLE 10

Chemical* and Normative Compositions of Two Denault Samples

	5-0	1-3
SiO_2	28.1	0.5
TiO_2	0.1	0.0
Al_2O_3	4.0	0.4
Fe_2O_3	0.2	0.2
FeO	1.51	0.17
MnO	0.0	0.0
MgO	13.7	21.0
CaO	18.9	29.6
Na_2O	0.1	0.1
K_2O	1.6	0.1
H_2O	0.53	0.18
P_2O_5	0.0	0.0
CO_2	30.20	46.30
C	0.0	0.22
Total	98.9	98.8
Dolomite	64.0	98.6
Quartz	21.6	--
Muscovite	11.8	0.7
Albite	0.8	0.5
Limonite	1.6	--
Sphene	0.2	--
Carbon	--	0.2

5-0. Laminated siliceous dolomite.

1-3. Stromatolitic dolomite.

*Analyst: K. Hoops, Geol. Surv. Canada

Carbon was separated from representative samples of the stromatolite types by digesting the samples in excess hydrochloric and hydrofluoric acids. X-ray diffraction of the residues thus obtained demonstrated the absence of crystalline structure. The residues oxidized readily at approximately 600° C. These observations indicate amorphous carbon rather than graphite.

Origin and Depositional Environment

The dolomitic composition of the Denault introduces the perennial dolomite problem, which has been well outlined by Van Tuyl (1914) and recently reviewed by Fairbridge (1957). The main theories of dolomite origin are:

1. Dolomitization of calcite or aragonite.
2. Primary precipitation of dolomite.

Van Tuyl (1914, p. 392) observed that dolomitization normally results in an increase of grain size. Long, in a study of the Conococheague limestone (1953, p. 107, p. 110), noted that dolomites of probable primary origin have a grain size of 1 to 7 microns whereas dolomites showing definite evidence of replacement origin have a grain size of 0.01 to 0.1 mm. According to Pettijohn (1957, p. 419), dolomitization also tends to destroy sedimentary textures and structures. In the Denault formation, grain size of less than 0.01 mm is common, and sedimentary textures and structures are remarkably well preserved. Coarse grained dolomite does occur in the massive beds, but has clearly formed by recrystallization of microcrystalline dolomite rather than by replacement of calcite or aragonite. Grain size, texture, and structure of the Denault therefore support primary deposition.

Precipitation of dolomite on the sea floor is unknown today (Fairbridge, 1957, p. 126). However, Chilingar (1956, p. 2262) caused precipitation of dolomite from sea water in the laboratory by saturating one litre sea water with respect of $MgCO_3$ and $CaCO_3$ and subjecting it to a CO_2 pressure of 4 atmospheres for 2 weeks. This experiment lends support to Strakhov's contention (1953) that a high CO_2 pressure of the

Precambrian atmosphere may have frequently resulted in direct precipitation of dolomite throughout much of Precambrian time. Primary precipitation of dolomite in evaporitic basins has also been proposed (Twenhofel, 1932, p. 339; Andrichuk, 1955; Edie, 1956). According to Hardley (1938, p. 1338) a mineral similar to dolomite is now precipitating in Great Salt Lake, Utah, and Strakov (1953) reported primary dolomite in Lake Balkash, Russia.

Because most of the Denault lithologies are clastic, earlier sedimentation to form source rocks is necessitated. Initial precipitation of calcite or aragonite may therefore have occurred in the source area, provided that dolomitization took place prior to the fragmentation and redeposition which formed the clastic dolomites. The writer has noted, however, the possibility of carbonate precipitation by the digitate and pisolitic stromatolites. These stromatolites are composed of finely crystalline dolomite, yet show extremely delicate primary structure. Direct precipitation of the dolomite is strongly suggested. Chave (1954, p. 281) has shown that some types of algae precipitate calcite or aragonite containing as much as 30 percent $MgCO_3$ in solid solution. Perhaps algae are capable of precipitating pure dolomite under conditions of increased CO_2 pressure or high salinity.

The columnar and undulatory stromatolites have been interpreted as sediment-binding algal structures. They are however, commonly associated with the pisolitic and digitate stromatolites (see columnar sections, pocket at back), and possibly the detritus which they trapped was derived by fragmentation and comminution of dolomite precipitated by digitate and pisolitic stromatolites. Thus, nearly half of the Denault may have formed as a direct result of algal precipitation.

The fine-grained dolomite of the laminated and massive beds may also have been derived by comminution of structures formed by lime-precipitating algae, but because these lithologies are non-carbonaceous and do not contain any recognizable stromatolite detritus, origin by inorganic precipitation is more likely. Influx of terrigenous detritus occurred during precipitation of the finely laminated beds, and slight variations in current velocity are indicated by the common occurrence of graded bedding of the waning current type. The dolomite sands of the massive beds may have formed by erosion of partially lithified muds or by aggregation of dolomite detritus.

All lithologies of the Denault formation show evidence of deposition in relatively shallow water. Because the Denault is more than 3000 feet thick, subsidence in pace with rate of sedimentation is therefore inferred. Dolomitic formations occupying similar stratigraphic positions occur in many parts of the Labrador Trough. Roscoe (1957, p. 7) and Fahrig (1955, p. 5) report stromatolitic dolomites, which probably correlate with the Denault, more than 100 miles north of Marion Lake. One hundred and fifty miles south of Marion Lake, Duffell (1959) mapped a marble formation which probably is the metamorphic equivalent of the Denault. The possible areal extent and great thickness of the Denault strongly suggest marine rather than lacustrine environment of deposition. The Green River formation of Wyoming, Colorado and Utah is an areally extensive continental deposit which may be more than 2000 feet thick (Bradley, 1929), but unlike the Denault, it contains abundant terrigenous sediments and only a few thin, relatively pure carbonate beds. Thick, widespread formations composed dominantly of carbonate rocks are abundant only in marine sequences (Barrell, 1906, p. 567).

Silicification of the Denault

Chert, chalcedony, and quartz are abundant in the Denault, particularly in the massive units. The chert occurs as nodules, most of which have the shape of oblate spheroids, flattened parallel to bedding. Boundaries are sharp and smoothly curving, and cross-sectional outlines are strongly elliptical, with axial ratios as high as 5:1. Nodules as large as 10 inches in diameter are common, but most are less than 4 inches in diameter. Several zones of abundant nodules can be recognized in the measured sections (pocket at back). Most of the nodules are white to gray, but in one zone (Unit 6) they are dark gray to black. Many of the nodules contain irregular remnant patches of dolomite, embayed by the chert, and some exhibit relict bedding continuous with that of the enclosing dolomite. The nodules are therefore interpreted as epigenetic replacement bodies. The nodules are commonly intersected by joints, which, as later noted, probably formed prior to folding. This, and the restriction of nodules to stratigraphic zones, strongly suggest diagenetic origin.

Bigge (1957), in considering origin of similar chert nodules concluded that the silica composing the nodules was derived from the host rock. This source would be adequate in the Denault, because all of the nodular zones contain abundant dispersed silica. However, the laminated dolomite, most siliceous of the rock types, contains very few nodules. Therefore, if the dolomites supplied the silica, the extent of nodule formation must have been controlled by factors other than the availability of dispersed silica.

The quartz and chalcedony form scattered lenticular pods, veinlets parallel to jointing, and irregular layers accentuating bedding and

stromatolite lamination (Plate 20). Much of this silica must have been introduced into voids, because crystals commonly line the walls of the occupied spaces, and many of the chalcedonic pods exhibit colloform banding. In places the silica infillings truncate sedimentary structures, suggesting solution origin for the voids. Soft, unconsolidated carbonate muds and sands would be incapable of supporting a network of voids, and therefore the solution and silicification probably occurred after lithification. However, many joints are unoccupied, suggesting that silicification ceased before all joints had developed. The veinlets parallel to jointing commonly cut the layers parallel to bedding, showing that the bedding-controlled silicification preceded at least some of the jointing-controlled silicification.

Some stromatolite zones are extensively silicified. The columnar and undulatory stromatolites are particularly susceptible. Sedimentary origin for the silica was at first suspected, but the following lines of evidence indicate that post-sedimentary origin is more likely:

1. Some highly silicified stromatolites grade laterally to stromatolites showing little or no silicification.

2. Quartzitic laminations commonly merge with joint-controlled veinlets showing no cross-cutting relationships.

3. Crystallographic c axes of quartz crystals are normal to the walls of the layers.

Silica in the silicified stromatolite zones composes as much as 60 percent of the stromatolites, but content of dispersed silica in non-silicified stromatolites in the same zones rarely exceeds 15 percent. This clearly indicates introduction of the silica rather than segregation of silica in the stromatolites.

PLATE 20

Silicified dolomite

A. Abundant quartz pods and stringers parallel to bedding in siliceous dolomite.

B. Extensively silicified columnar stromatolite unit.

Wishart Formation

The Wishart formation overlies the Denault with slight unconformity. The sharp contact is exposed in several outcrops on the south shore of Marion Lake. Three measured section show that the formation is approximately 300 feet thick, and is composed of 45% sandstone, 25% ankeritic siltstone, 15% siltstone, 10% cherty siltstone, and 5% conglomerate. A generalized section is shown in Table 11.

TABLE 11

Generalized Section of the Wishart Formation

Unit	Approximate Thickness (feet)
Siltstone	10
Orthoquartzite	30
Siltstone	10
Ankeritic siltstone	30
Cherty siltstone	15
Lenses of ankeritic conglomerate	0-5
Upper chert pebble conglomerate	10-15
Cherty siltstone	15
Ankeritic siltstone	45
Orthoquartzite	100
Lower chert pebble conglomerate	0-10
Siltstone	20
Total	295-315

Sandstone

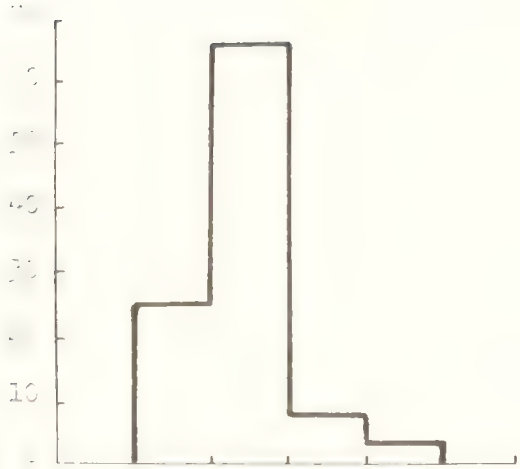
Description

The sandstones are tough, massive, medium-grained rocks, which are white or light gray on fresh and weathered surfaces. They are very well-sorted, chert- and quartz-cemented orthoquartzites (Table 12). Well-rounded, highly spherical quartz and chert grains are the only framework minerals (Plate 21). Overgrowth rims in crystallographic continuity with the detrital quartz grains can be seen in some specimens. The quartz grains contain planes of tiny opaque inclusions and show strong undulatory extinction. Because the traces of some inclusion planes extend across several adjacent grains, the inclusions must have originated, at least in part, during metamorphism of the Knob Lake group. Chert is the cement in some of the sandstones, but most of the rocks are welded together by mutual interpenetrations of grain boundaries. Flakes of 2M muscovite, less than 0.01 mm long, are concentrated along the finely sutured boundaries of most sand grains. Colorless zircon and greenish blue tourmaline are accessory detrital minerals in the orthoquartzites. They occur as well-rounded detrital grains less than 0.1 mm in diameter. Magnetite, ilmenite, leucoxene, and limonite are the major opaque minerals.

Size frequency distributions for two Wishart sandstones were calculated from thin section measurements (Appendices A and B). Histograms of the distributions are shown in Fig. 8. For both samples the mode is in the 0.25-0.5 mm class, and the size range is restricted to four classes. The distributions are skewed towards the coarser grades.

Fig. 8

Size histograms for two Wishart sandstones, showing the estimated volume percent distribution of detrital grains.


10
1
Particle Size, ϕ Units

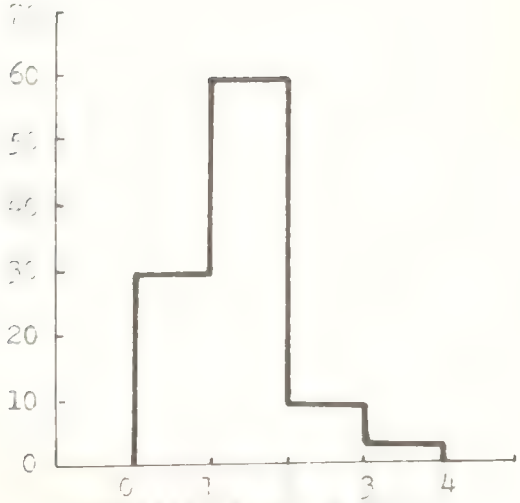
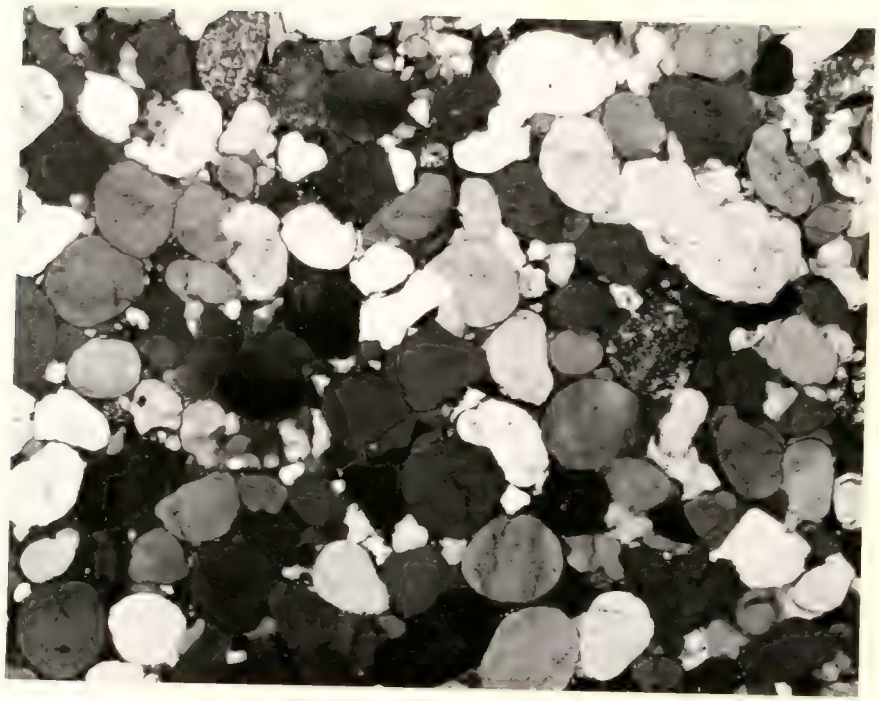

60
20
10
0
0
3
4
Particle Size, ϕ Units

PLATE 21

Photomicrographs of Wishart sandstone

A. Thin section of upper orthoquartzite unit. Note excellent rounding and sphericity of quartz and chert grains. Crossed nicols, X30.

B. Thin section of lower orthoquartzite unit. Note abundant mutual interpenetrations of the quartz grains. Crossed nicols, X30.



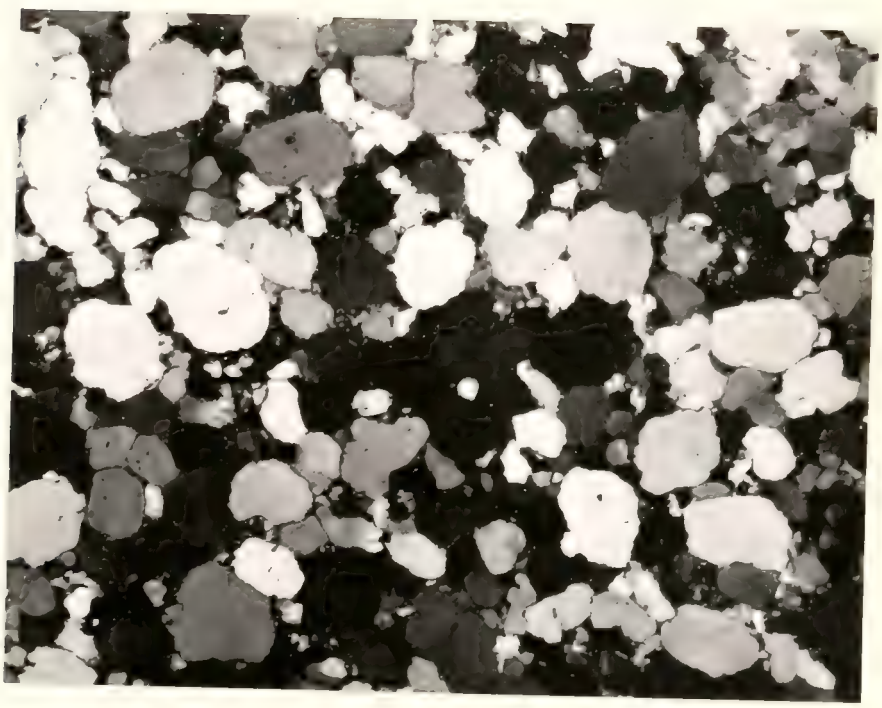


TABLE 12

Modal Composition of Some Wishart Sandstones

(Volume percent, based on 800 points for each specimen)

Specimen No.	Unit	Quartz Grains	Chert Grains	Chert Cement	Muscovite	Opaques	Tourmaline	Zircon	Average Roundness	Average Sphericity	Median Diameter	Sorting Coefficient S_o	Classification (Pettijohn, 1957)
58-6-3	3	90.5	—	2.0	2.9	4.6	*	*	0.7	0.8	0.67	1.32	Orthoquartzite
58-6-4	3	91.7	—	1.4	5.1	1.8	*	*	0.6	0.8	0.64	1.18	Orthoquartzite
58-6-16	13	85.2	2.7	10.7	*	1.4	*	*	0.9	0.9	0.62	1.15	Orthoquartzite

*Less than 0.2 percent.

Origin and Depositional Environment

Sandstones of the Wishart are very mature orthoquartzites which have probably passed through several cycles of sedimentation. The well-rounded sands may have undergone extensive wind transport prior to deposition because, according to Twenhofel (1945, p. 59), wind is more effective than water in the rounding of sand grains. Dake (1921, p. 223) and Thiel (1935, p. 612) in detailed studies of the well known St. Peter sandstone, which shows rounding equal to that of the Wishart, suggested some aeolian abrasion prior to deposition to account for the excellent rounding.

Excellent sorting of the Wishart sands strongly suggests deposition in a littoral or aeolian environment. Many modern aeolian and beach sands (Udden, 1914, p. 703; 714-719), show similar distribution. The sands were probably derived from parts of the Marion Lake formation which were exposed either by local upwarp and erosion, or by virtue of non-deposition of the Dersult and Keco formations in the source areas. The sands exhibiting the best rounding and sorting occur near the top of the Wishart, and possibly were derived by reworking of older Wishart sand units. It could be argued that reworking should result in reduction of size, yet size distributions of the Marion Lake sandstones do not include larger size grades than distributions of the Wishart sandstones. In fact, some samples of the Marion Lake sandstones are finer grained than the Wishart samples. However, it is not unreasonable to postulate source rocks of Marion Lake age which were coarse-grained lateral equivalents of the Marion Lake sandstones exposed in the map-area.

Conglomerate

Description

The lower conglomerate is best exposed on the south shore of the northwest arm of Marion Lake. It forms lenticular beds, less than 10 feet thick, containing subangular to rounded fragments of greenish-gray chert, white vein quartz, and gray to white arkosic quartzite. The fragments, 0.5 to 3 cm in diameter, compose approximately 30 percent of the rock, and are set in a medium-grained protoquartzite matrix (Plate 22A). Sand grains of the matrix are very well rounded and highly spherical. Most of the grains are quartz, but there are a few sand-size fragments of chert. The quartz and chert sand grains are cemented by a very fine-grained muscovite intergrowth containing irregular patches of chert. Pyrite and leucoxene are the most abundant opaque minerals. Detrital zircon grains and euhedral crystals of apatite, less than 0.01 mm in diameter, are minor accessory minerals.

The upper conglomerate crops out on the small island in the northwest arm of Marion Lake, and on the north shore of the lake, one mile southwest of the island. In both places, marked graded bedding can be seen. Angular and rounded pebbles of jasper and chert, as much as 5 cm in diameter, occur near the base of the conglomerate, but only scattered granules occur near the top. The pebble-sized fragments of chert and jasper compose less than 40 percent of the gray-black, to greenish-black, rusty weathering conglomerate (Plate 22B). The matrix consists of intimately intergrown chert, stilpnomelane, ankerite, and an acicular mineral which may be minnesotaite. The latter mineral has the pleochroic formula: X = pale yellow, Y = Z = light green, shows moderate birefringence,

PLATE 22

Wishart conglomerate

A. Photomicrograph of lower conglomerate showing marginal replacement of chert pebbles (light-colored) by fine-grained ankerite. Note euhedral authigenic siderite crystal (extensively altered to iron oxides) near center of photograph. Ordinary light, X2.

B. Polished surface of upper conglomerate showing jasper and chert pebbles in a matrix rich in iron silicates. Note colloform banding in pebble, lower left corner. (X2).

is length slow, and has indices $N_x = 1.60$; $N_z = 1.64$. It shows parallel extinction and 2V is negative and small. These data match the optical data for minnesotaite (Winchell, 1951, p. 365), but do not eliminate some of the chlorites, particularly aphrosiderite (Winchell, 1951, p. 384). Pyrite, hematite, limonite, amorphous carbon, and apatite are minor constituents.

On the east side of the island, several small outcrops of cream-buff, rusty-weathering conglomerate containing as much as 50 percent ankerite occur at the top of the upper conglomerate. Well-rounded sand grains of quartz and chert compose more than 15 percent of the rock, and light gray subangular chert pebbles compose most of the remaining 35 percent. Opaque minerals are virtually absent, but apatite is abundant.

Origin and Depositional Environment

The lower Wishart conglomerate resembles the Marion Lake conglomerates in composition, lenticularity, and lack of sorting, and, like the Marion Lake conglomerates, is perhaps best interpreted as a fluviatile deposit. Roundness of sand grains in the matrix was probably inherited from a previous cycle. The abundant quartzite pebbles support this view, and their arkosic composition suggests the Marion Lake formation as the probable source.

Notable in the upper conglomerate is the exclusive occurrence of chert and jasper fragments in the framework. Most of these fragments show no rounding, indicating a relatively short history of transport. The ironstone matrix shows no evidence of clastic origin, and because the chert pebbles form an open framework, the matrix clearly cannot have originated by infiltration. The matrix is therefore interpreted as

chemically precipitated sediment. Intimate association of fine grained chemical and coarse clastic components is anomalous, but because the graded conglomerate is a single occurrence associated with clearly shallow water sandstones, it can perhaps be attributed to near-shore slumping of lithified chert into an environment of ironstone deposition.

Siltstone

Description

The siltstones and cherty siltstones of the Wishart formation are closely associated rock types. Both are tough, thinly bedded, and appear light gray to pale greenish buff on both fresh and weathered surfaces. The cherty siltstones tend to break with a conchoidal fracture. Bedding in the siltstones is marked by slight differences in grain size, and in the cherty siltstones it is accentuated by alternation of chert-rich and silt-rich laminations (Plate 23A). Most of the silt grains are well-rounded, less than 0.05 mm in diameter, and exhibit undulatory extinction. Muscovite, in amounts up to 10 percent, is present in all specimens. The muscovite occurs as shreds less than 0.01 mm long, parallel to the bedding. Rhombic crystals of siderite, extensively pseudomorphed by limonite, are commonly strung out parallel to bedding. The rhombs average 0.05 mm in diameter, enclose clastic quartz grains, and show distinct euhedral outlines (Plate 23B).

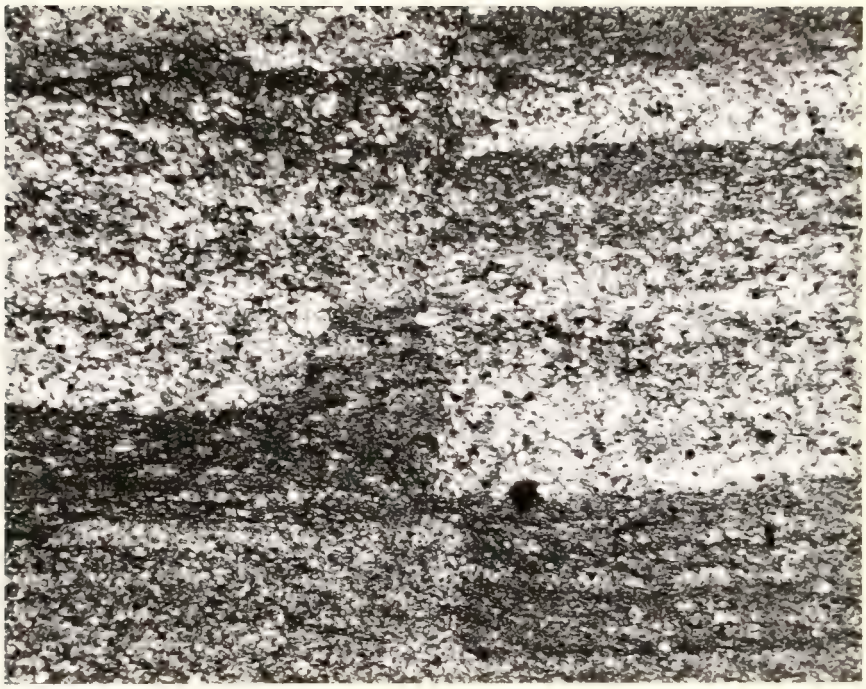
The ankeritic siltstones are characterized by a powdery, yellowish-brown limonitic rind on weathered surfaces. Fresh specimens are tough, massive, and light gray in color. Slight effervescence is produced by cold 20 percent hydrochloric acid. Indices of refraction for the carbonate mineral in several samples were determined as N_O = 1.72-1.75 and

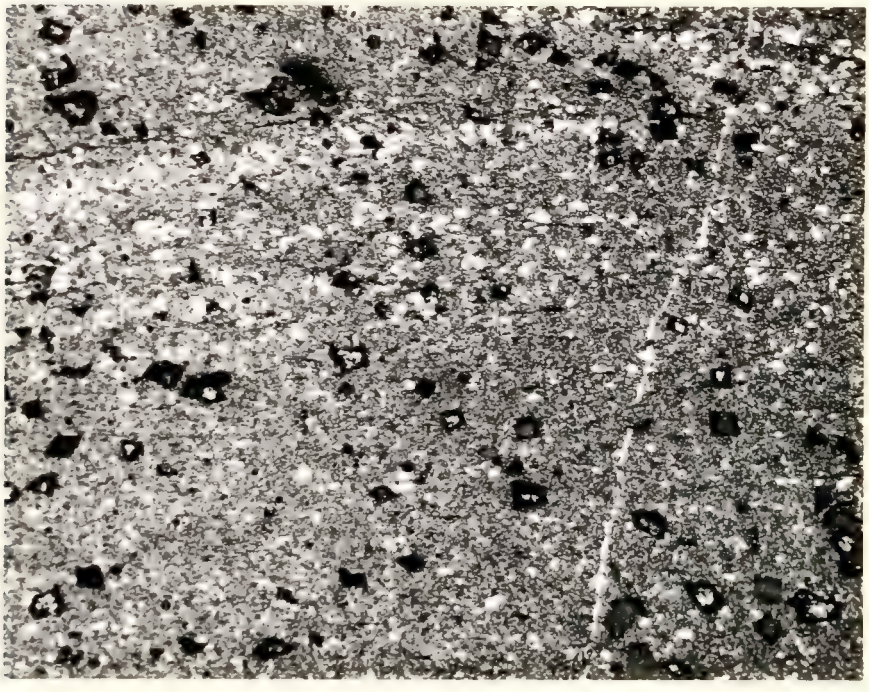
PLATE 23

Photomicrographs of Wishart cherty siltstone

A. Thin section of cherty siltstone showing offset along a microfracture. Ordinary light, X30.

B. Thin section of cherty siltstone containing limonite pseudomorphs after authigenic siderite euhedra. Crossed nicols, X40.





N_E = 1.52-1.54, indicating ferrodolomite to ankerite composition (Winchell, 1951, p. 114). Dolomite structure was verified by x-ray diffraction. The ankerite grains, which compose approximately 30 percent of the ankeritic siltstones are neutral to pale brown in thin section, and lack lamellar twinning. The ankerite grains are anhedral, and most are less than 0.01 mm in diameter. They are intimately intergrown with chert, forming a fine-grained matrix for the quartz silt (Plate 24A). Angular to poorly rounded fragments of collophane are present in some specimens (Plate 24B). The fragments, which are up to 0.5 mm in diameter, are altered in part to clusters of clear hexagonal crystals less than 1 micron in diameter. The crystalline alteration is probably either dahllite or apatite. Muscovite flakes less than 0.01 mm long are minor constituents in most specimens.

Origin and Depositional Environment

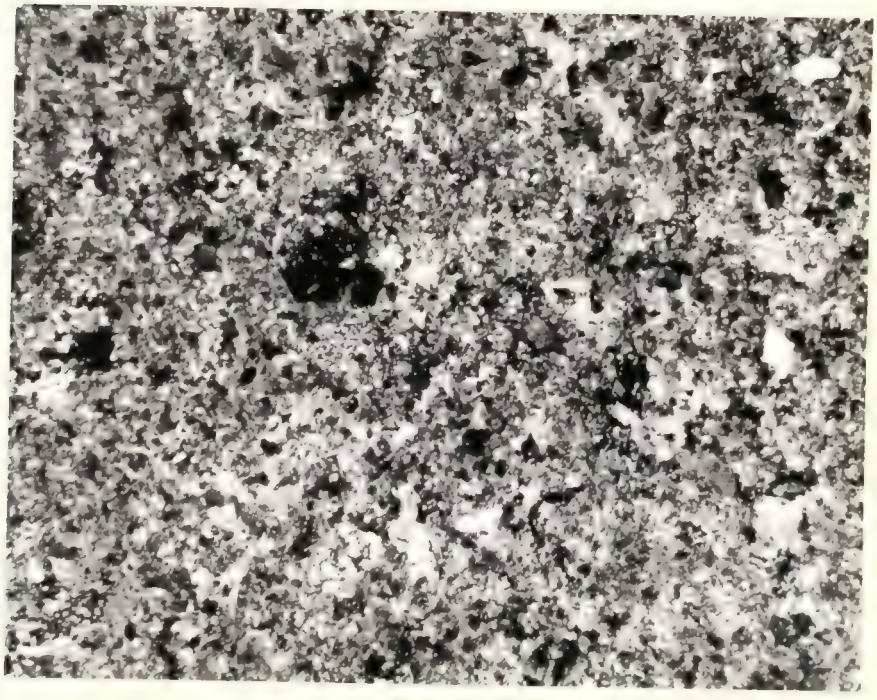
Fine lamination of the Wishart siltstones indicates deposition below wave base, but because of association with the well-sorted sandstones of probable shallow water or aeolian origin, no great depth of water is inferred. The fine-grain size and intimate intergrowth of the ankerite and chert suggests chemical origin for these minerals. The silt grains, however, are clearly detrital, and therefore clastic as well as chemical sedimentation must have taken place. Krumbein and Garrels (1952) have shown the importance of Eh and pH in precipitation of chemical sediments. According to their studies, siderite will precipitate in a slightly reducing environment with a pH range of 7.0 to 8.0. No data are given for ankerite, but presumably conditions for its precipitation would be similar to those required for siderite precipitation. Krumbein and

PLATE 24

Photomicrographs of Wishart ankeritic siltstone

A. Fine-grained ankeritic siltstones showing detrital quartz silt in a fine-grained matrix of chert and ankerite. Ordinary light, X20.

B. Thin section showing two angular detrital fragments of collophane (dark patches near center and at left) in an ankeritic siltstone. Ordinary light, X45.



Garrels (1952, p. 19) suggest that chert is Eh- and pH- independent. Correns (1950) considers organic activity essential to primary precipitation of chert, and Bramlette (1946, p. 42) has shown that the thick Monterey formation of California consists mostly of altered diatomaceous deposits, but other workers (Van Hise and Leith, 1911, p. 516; Davis, 1918) hold that inorganic precipitation is possible in environments to which silica is contributed by vulcanism. Taliaferro (1933) suggests that volcanic activity may be necessary to sufficiently increase the silica content of sea water to allow abundant organic precipitation of silica. Evidence of vulcanism contemporaneous with Wishart sedimentation cannot be shown for the Marion Lake area, but because volcanic rocks are abundant in the Labrador Trough, silica enrichment due to vulcanism must be considered possible.

The siderite rhombs in the siltstones must be either authigenic or metamorphic because they are euhedral and commonly enclose grains of quartz silt. Alteration to limonite is the result of weathering.

The occurrence of collophane in the Wishart rocks may well be unique in the Precambrian record. Because the collophane is fragmental, phosphorite source beds are probably present elsewhere in the Labrador Trough. The fragments could also represent the phosphatized hard parts of primitive invertebrate life forms. Mansfield (1927, p. 366) interprets phosphatic sediments of the well-known Permian Phosphoria formation as formed by "... biochemical and physical agencies from phosphatic solutions or colloids on the sea bottom." Pettijohn (1957, p. 475) notes that phosphorites are deposited typically in marine environments under anaerobic conditions.

Menihek Formation

A thick sequence of slate, shale, argillite, and quartzite lies stratigraphically above the Wishart formation. Frarey mapped the westward extention of these rocks as the Howse group, but he suggested the possibility of their equivalence to the Menihek formation of the type section at Knob Lake (Frarey, 1954).

Within the Marion Lake area, there are several places along the north shore of the northwest arm of Marion Lake where slate crops out less than 100 feet stratigraphically above the Wishart. The contact is not exposed, but bedding attitudes strongly suggest conformity. Further, large-scale folds in the slate belt, accentuated by conformable basic sills, correspond closely to the large-scale folds in the Denault formation. The rocks above the Wishart are therefore here assigned to the Menihek formation.

Description

Slate and shale compose more than 90 percent of the Menihek. These rocks are greenish gray to grayish black on fresh surfaces, and yellowish brown to light gray on weathered surfaces. Bedding is commonly obscured by cleavage, but where visible, is seen to consist of tabular laminations less than 1 mm thick. Gray, silty quartzitic layers are alternated with dark, finer grained micaceous layers. The silty layers commonly show delicate cross-lamination. Detrital quartz grains, up to 0.03 mm in diameter, can be seen in thin section, but most of the minerals cannot be accurately identified by means of the microscope. X-ray analyses reveal that quartz, albite, chlorite, and 2M muscovite are the major

minerals. Small amounts of pyrite, dolomite, carbon, and iron silicates are present in most specimens. Hematite and magnetite are abundant in beds near the base of the Menihek formation.

In Table 13, chemical composition of a sample from the lower part of the Menihek formation in the Marion Lake area is compared with compositions of two Ruth samples and two Menihek samples from the Knob Lake area (Gross, 1951). The slate from the Marion Lake area more closely resembles the Ruth slates in having low silica and high total iron contents, but differs from them in having higher soda, lime and magnesia contents. The titania content is unusually high in (A) and (B). Specimens taken from the upper part of the Menihek in the Marion Lake area are more siliceous than the slate submitted for analysis (A), and x-ray analyses do not reveal the presence of abnormal amounts of iron oxides. The chemical composition of the upper Menihek in the Marion Lake area would therefore more closely approximate in composition the type Menihek at Knob Lake (D, E). This substantiates the suggestion that the rocks mapped as "Menihek" in the Marion Lake correlate with the Ruth-Sokoman-Menihek sequence of the type locality.

Argillites of the Menihek are similar in composition to the slates and shales, but lack well-developed lamination. Fine- to medium-grained quartzites form widely scattered lenticular beds in the Menihek. Recrystallization has obscured most detrital outlines, but where visible, the grains are well-rounded and uniform in size. The quartzites are massive gray rocks which commonly show cross-bedding. They are composed mostly of quartz and minor muscovite and chlorite. The micaceous minerals commonly occur in clusters which may represent distorted shale fragments. The original sandstones were probably protoquartzites, according to Pettijohn's classification (1957, p. 291).

TABLE 13

Chemical Composition of Menihek and Ruth Slates

	A	B	C	D	E
SiO_2	53.9	52.35	54.84	66.88	71.22
TiO_2	2.3	1.86	0.42	0.82	0.48
Al_2O_3	14.9	11.56	10.62	14.09	13.04
Fe_2O_3	9.0	13.09	13.39	1.96	1.39
FeO	4.57	3.98	7.50	1.41	3.09
MnO	0.1	0.15	- -	- -	- -
MgO	4.3	- -	1.79	1.52	2.36
CaO	1.3	0.34	0.52	0.53	0.42
Na_2O	2.1	1.32	0.13	0.78	1.14
K_2O	2.8	4.54	3.06	4.15	2.82
H_2O +	3.66	7.34	4.46	1.94	2.84
H_2O -		1.33	0.97	1.28	0.42
CO_2	0.00	- -	- -	- -	- -
P_2O_5	0.2	0.25	0.11	0.10	0.13
C	- -	1.78	2.18	4.24	0.86
Total	99.1	99.89	99.99	99.70	100.21

A. Sample of the lower part of the Menihek formation, Marion Lake area. Sample 6-19.
Analyst: G. Bender, Geol. Surv. Canada.

B, C. Samples of the Ruth slate, Knob Lake area.
Analyst: W. H. Herdsman, Glasgow.

D, E. Samples of the Menihek slate, Knob Lake area.
Analyst: W. H. Herdsman, Glasgow.

Origin and Depositional Environment

Delicate lamination and fineness of grain in the Menihek shales and slates indicate a relatively quiet water environment of deposition. The quartzites, however, are well-sorted and cross-bedded, suggesting deposition in a turbulent environment. Sedimentation may have been controlled by water depth, sand being deposited when wave base was depressed to the depositional interface. Introduction of the sandy units to a deep basin by turbidity flow is another possibility, but maturity of the quartzites and lack of graded bedding stand against such origin.

High titania content of the Menihek rocks may have resulted from unusual concentration of titanium-bearing minerals in the sedimentary environment, addition of titania during intrusion of the sills, or metasomatic transfer during regional metamorphism. The high titania content could also reflect contemporaneous vulcanism with release of titania to the waters in the basin of Menihek deposition. Because the lower beds of the Menihek are iron-rich, a relationship may well exist between vulcanism and Menihek sedimentation. Van Hise and Leith (1911, p. 513-516) suggested that contemporaneous basic eruptive rocks probably yielded the iron in the iron-bearing sediments of the Lake Superior iron region through one, or a combination of, the following processes:

1. Transfer in hot magmatic solutions migrating from the eruptive lava during it solidification.
2. Action of meteoric waters on subaerial flows.
3. Transfer by direct reaction of the hot magma with sea water.

James (1951) however, has advanced the theory that climate is the important control of iron-sedimentation. James notes that humid-tropical climate in modern environments is favorable for iron-enrichment, and he

stated (1951, p. 264): "It is the writer's opinion that climate alone, with its attendant effects on weathering and stream load, is eminently qualified to furnish the transcending control for an interval of iron-rich sedimentation."

MURDOCK GROUP

The Murdock group is separated from the Knob Lake group by a northwesterly trending fault, and crops out east of the fault in a belt less than one mile wide. Chlorite-biotite-quartz schists and chlorite-albite-actinolite schists compose most of the Murdock, but massive intrusive rocks are locally abundant.

Description

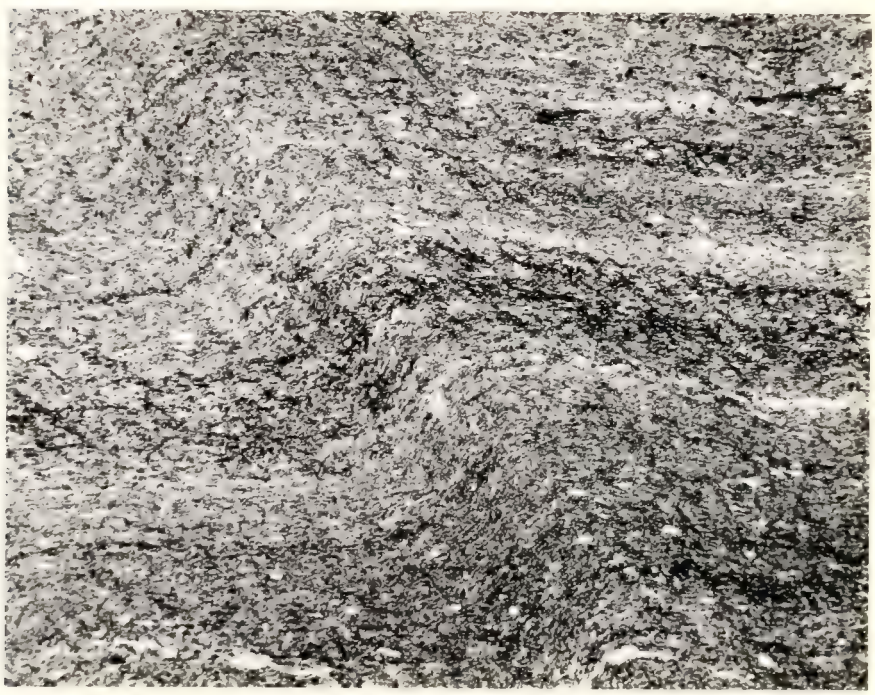
The fine- to medium-grained chlorite-biotite-quartz schists are dark green to greenish gray on fresh surfaces, and rusty brown to pale green on weathered surfaces. Some outcrops show relic bedding averaging 0.5 to 3 mm thick, but it is commonly obscured by cleavage. Typical specimens contain approximately 35% quartz, 30% biotite, 25% chlorite, 5% muscovite and 5% albite. Because untwinned albite may be mistaken for quartz, the feldspar to quartz ratio may be somewhat higher. Minor actinolite, magnetite, and granular, colorless to pale yellow epidote are present in some thin sections, and one specimen contains more than 25 percent carbonate (Plate 25). The quartz and feldspar grains are equant and less than 0.1 mm in diameter. The biotite occurs in shreds up to 0.2 mm long, and exhibits yellowish brown to greenish brown pleochroism. The chlorite laths are less than 0.1 mm long, and are pleochroic from neutral to pale green.

PLATE 25

Photomicrographs of Murdock schist

A. Thin section showing intricate crenulated flow cleavage (horizontal) in a specimen of Murdock schist. Note faint, steeply dipping fracture cleavage. Composed essentially of chlorite (light gray), biotite (dark gray), carbonate (white), and magnetite (black). Ordinary light, X20.

B. Thin section showing folded flow cleavage (horizontal). Note the fracture cleavage parallel to the axial plane of the small fold. Ordinary light, X20.



The chlorite-albite-actinolite schists have fine-grained, greenish gray fresh surfaces and gray to brown weathered surfaces. Some outcrops show relic pillows and pyroclastic textures. Typical specimens contain approximately 45% actinolite, 30% albite, 15% chlorite, 10% clinozoisite, and minor sphene, magnetite, and leucoxene. Both actinolite and chlorite are pale green and slightly pleochroic. The albite occurs as clear, anhedral grains, and the clinozoisite forms granular clusters. Sphene, magnetite, and leucoxene compose as much as 10 percent of some specimens (Plate 25B).

The massive rocks are fine- to medium-grained, and are composed mostly of saussuritized plagioclase and uralitized pyroxene. They are tough, greenish gray, rusty weathering rocks which closely resemble the gabbros associated with the Knob Lake and Doublet groups.

Origin

Relic bedding in the chlorite-biotite-quartz schists indicates a sedimentary origin. The mineralogy of these rocks suggests derivation from immature pelitic sediments or tuffs. Relic pillow structures and pyroclastic textures in the chlorite-albite-actinolite schists demonstrate derivation from volcanic rocks, and the mineralogy suggests an original basaltic composition for the altered rocks.

Frarey (1954) reported outcrops of conglomerate, containing pebbles of chert and iron formation, at the base of the Murdock group in the Willbob area. He cited this occurrence as evidence for an unconformity between the Knob Lake and Murdock groups. The Murdock conglomerate was not observed in the Marion Lake area, but because of the relationship observed by Frarey, the Murdock group is here regarded as younger than the Menihek.

Similarity in mineralogy of Murdock and Doublet rocks (Table 17) suggests that Murdock rocks might be schistose Doublet equivalents, but attitude determinations in the Willbob area (Frarey, 1954) show that the Murdock is a monoclinal sequence facing northeast. Because lowermost Doublet volcanic rocks in the Marion Lake area also face northeast, equivalence of Doublet and Murdock groups would require repetition by major faults for which there is no evidence.

DOUBLET GROUP

Rocks of the Doublet group underlie about one-half of the Marion Lake area. A sedimentary sequence less than 3000 feet thick comprises the lower part of the group; the upper part is a volcanic sequence with a thickness of more than 10,000 feet. The Doublet volcanic rocks occupy a belt which probably terminates just south of Roymart Lake, but which extends more than 80 miles northwest of the map-area. Similar volcanic rocks have been mapped in the vicinity of the Larch River, 250 miles northwest of Marion Lake (Sauvé, 1957).

Doublet Sedimentary Rocks

Beds of quartzite, slate, and argillite constitute the basal sequence of the Doublet group. These rocks are best exposed north of Link and Moss Lakes, and on the limbs of the folds which flank Lac La Touche. A few thin slaty and tuffaceous beds are intercalated with the pillow lavas, but are not shown on the map.

Quartzite

Description

Most of the quartzites are massive, fine- to coarse-grained, light gray to greenish gray rocks which weather white to rusty brown. Bedding is commonly obscure or absent, but in some of the finer-grained quartzites, black shaly partings mark distinct planar beds ranging from 1 to 5 mm in thickness. Planar cross-beds as much as 2 cm thick occur in some of the quartzites, and delicate undulatory structures, with an amplitude of 1 to 2 mm and wave lengths of 5 to 15 mm, were observed in the fine-grained quartzites which crop out one-half mile north of Moss Lake. The undulatory structures occur on bedding surfaces and may be ripple marks. They are very small-scale features, but Bucher (1919, p. 158, Table III) recorded modern ripple marks with amplitudes as little as 1 mm and wave lengths of 11 mm.

Quartz sand grains compose 55 to 65 percent of the quartzites. The grains are smoky, pale blue, or clear and glassy, and few are larger than 1 mm in diameter. Boundaries are delicately sutured due to mutual interpenetration of adjacent grains, but detrital outlines, marked by dusty rings within the serrated margins, can commonly be seen. Some grains exhibit high sphericity, but most are ovoid or lenticular; a few have length to width ratios of more than 3:1 (Plate 26). Slightly clouded grains of detrital plagioclase (mostly oligoclase) compose 2 to 5 percent of the quartzites, and a few detrital microcline grains are commonly present.

Chlorite and biotite, the most abundant micaceous minerals, occur as shreds less than 0.1 mm long, strung out along the margins of

PLATE 26

Photomicrographs of Doublet quartzite

A. Thin section of typical Doublet quartzite showing poorly sorted, angular quartz grains. Note large rounded quartzite grain. Crossed nicols, X20.

B. Thin section of arkosic Doublet quartzite. Note strong undulatory extinction of the quartz grains. Crossed nicols, X20.

the detrital grains. The chlorite is pleochroic from light green to olive green, and the biotite is pleochroic in reddish- or greenish-browns. Pyrite, epidote muscovite, leucoxene, iron oxides, and small, well-rounded zircons are accessory minerals. The rock fragments include fine-grained quartzite, chert, and chloritic shale. The shale fragments are commonly smeared out along cleavages. The content of quartzite fragments shown in Table 14 is probably too low, because much of the fine-grained quartz may have been derived from quartzite grains which were crushed during metamorphism. The strong undulatory extinction exhibited by the quartz, and the numerous fractures in the rocks support this view.

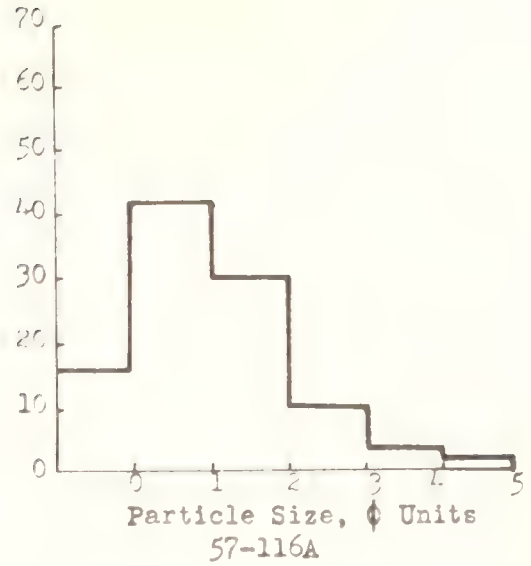
The modal composition of three quartzites are shown in Table 14. According to Pettijohn's classification (1957, p. 291), the rocks are lithic sandstones. The chemical composition of a Doublet quartzite is shown in Table 15. As in lithic sandstones from other areas (Pettijohn, 1957, p. 319), magnesia exceeds lime and soda exceeds potash. The sample has the following normative composition:

Quartz	68.7	Muscovite	1.6
Oligoclase	17.3	Sphene	0.4
Chlorite	11.8	Calcite	0.2

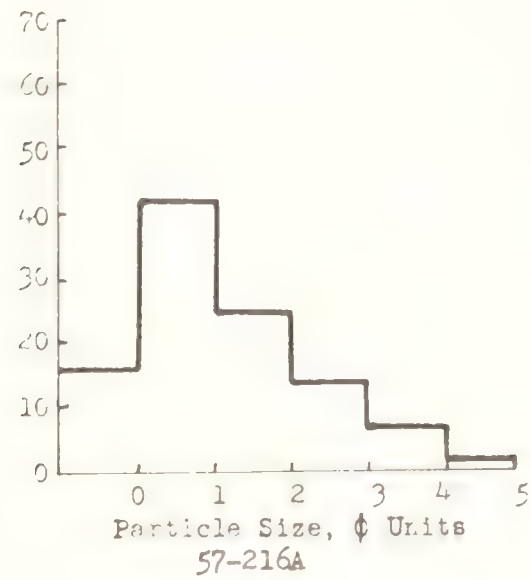
Histograms of size frequency distributions, calculated from thin section measurements (Appendices A and B), are shown for two Doublet quartzites in Fig. 9. The samples have coarse sand modal classes (0.5 - 1 mm), show fair sorting, and are skewed towards the coarse size grades. Because there is evidence of crushing and recrystallization in the samples, the original distributions probably included coarser grades.

Fig. 9

Size histograms for two Doublet quartzites, showing the estimated volume percent distribution of detrital grains.



57-116A



57-216A

TABLE 14

Modal Composition of Some Doublet Quartzites

(Volume percent, based on 600 points for each specimen)

Specimen No.	Quartz Grains	Quartzite Fragments	Chert Grains	Shale Fragments	Plagioclase	Microcline	Chlorite	Biotite	Muscovite	Zircon	Opaques	Epidote	Average Roundness	Average Sphericity	Median Diameter	S_o, Sorting Coefficient
57-116A	60.1	14.5	2.7	1.0	9.2	2.1	3.7	6.1	*	*	*	*	0.5	0.6	0.82	1.76
57-210A	55.1	15.2	—	1.9	13.8	—	10.9	—	1.7	*	0.4	*	0.6	0.5	0.37	1.33
57-216A	62.9	15.3	—	—	12.7	—	3.8	4.7	0.4	*	0.2	*	0.5	0.6	0.80	1.58

*Less than 0.2 percent.

TABLE 15

Chemical Composition of a Doublet Quartzite*

SiO_2	83.8	CaO	0.9
TiO_2	0.2	Na_2O	2.0
Al_2O_3	6.5	K_2O	0.2
Fe_2O_3	0.7	H_2O	1.22
FeO	3.25	P_2O_5	0.0
MnO	0.0	CO_2	0.10
MgO	1.5	Total	100.4

*Sample 57-210A

Analyst: G. Bender, Geol. Surv. Canada

Origin and Depositional Environment

Fair sorting, cross-bedding, and possible ripple marks of the Doublet quartzites indicate shallow water deposition. Similar lithic sandstones are interpreted as continental or shallow marine deposits (Krynine, 1940; Siever, 1949; Pelletier, 1958). According to Pettijohn (1957, p. 62), lithic sandstones belong to the realm of paralic sedimentation, which implies intense terrigenous alluviation. The mineral and chemical compositions of the Doublet quartzites reflect immaturity, in keeping with this concept. They are therefore interpreted as derivatives of sands which were subjected to a short but vigorous transport and deposited in a fluviatile, lacustrine, or shallow marine environment.

Slate and Argillite

Description

The Doublet slates are sooty black, grey to rusty weathering aphanitic rocks. Paper-thin beds are visible on the weathered surfaces of some specimens, but cleavage commonly masks the bedding. Only quartz, which occurs as grains less than 0.01 mm in diameter, can be definitely identified in thin section. Other mineral grains are only a few microns in diameter, and are felted together in an extremely fine-grained intergrowth. X-ray diffraction patterns of several specimens show the presence of abundant quartz and 2M muscovite. Minor chlorite, plagioclase and iron oxides were detected in all specimens. The chemical composition of a typical Doublet slate is shown in Table 16. The sample has a normal silica content, but is notably high in ferrous iron, and low in lime. The sample has the following normative composition:

Muscovite	31.6	Sphene	.3
Quartz	30.4	Ilmenite	.4
Chlorite	16.9	Calcite	.2
Albite	10.2	Carbon	.2
Iron Oxide	9.8		

Origin and Depositional Environment

The fine grain size and thin lamination of the Doublet slaty rocks indicate relatively quiet water deposition. Because they are interbedded with lithic sandstones of fluviatile, lacustrine, or shallow water marine deposits, the slates probably formed in similar environments during periods of quiescence.

TABLE 16

Chemical Composition of a Doublet Slate*

SiO_2	57.6	CaO	0.2
TiO_2	0.4	Na_2O	1.2
Al_2O_3	15.2	K_2O	3.8
Fe_2O_3	1.2	H_2O	4.22
FeO	13.48	P_2O_5	0.0
MnO	0.0	CO_2	0.09
MgO	3.2	C	0.23
		Total	100.0

*Sample 57-119A

Analyst: K. Hoops, Geol. Surv. Canada

Doublet Volcanic Rocks

Description

Typical volcanic rocks of the Doublet group are greenish gray and weather light gray to pale yellowish green. They are very fine-grained, tough, and even-textured. Many specimens show faint to well-developed flow cleavage, and some exhibit a prominent fracture cleavage.

Pillows are present in more than 80 percent of the mapped outcrops. Most pillows are less than 6 feet long, and many are less than 3 feet long, but several more than 20 feet long were observed north of Ryan Lake. According to Wilson's classification (1941) most of the pillows are balloon- and loaf-shaped, but matress- and bun-shaped pillows are abundant. The pillows fit closely together and the shapes commonly provide

a reliable indication of flow tops (Plate 27A). Most pillow selvages are 1 to 3 cm thick, and are commonly darker, but weather lighter, than the pillow interiors. Cavities in the upper portions of pillows also provide a top and bottom criterion (Plate 27B), and proved to be an invaluable aid to mapping. Sauvé (1957, p. 44-46) reviewed the theories of origin for pillow cavities and concluded that the pillow voids can form by contraction during cooling, sagging of the lower portions of the pillows, agglomeration of gas bubbles, capture of extraneous water, the draining out of liquid lava, or some combination of these processes. Most cavities are empty, but in places where quartz veins are abundant (particularly north of Moss Lake and west of Gauthier Lake), the cavities are commonly occupied by quartz. Quartz and minor chert also occur in pillow interspaces.

Few flow thicknesses can be accurately determined because the scoriaceous tops weather more rapidly than the central and basal portions of the flows, forming valleys identical in appearance to valleys controlled by strike joints. On several hill-tops in the northern part of the map-area, however, complete cross-sections of flows, 50 to 230 feet thick, are exposed. Three of these flows were traced on the ground for distance of more than 3 miles, and in the interval were seen to maintain very uniform thicknesses. Flow tops marked by parallel valleys can be traced for even greater distances on aerial photographs.

Porphyritic pillow lavas crop out on the north shore of Lac Gauthier, on the east shore of Lac Cherny, and in several places north of Lac La Touche. Some porphyritic flows can be traced around the noses of folds (see map, pocket at back). Glomeroporphyritic clots of altered plagioclase, as much as 3 cm in diameter, are scattered through the

PLATE 27

Pillowed Doublet metabasalt

A. Shore line exposure of Doublet metabasalt showing well-developed pillow structure. Flow is right side up and dips gently away from the observer.

B. Cavity in the upper portions of a pillow. Note glomeroporphyritic clots of plagioclase, and well-developed selvage.

porphyritic rocks. The clots are evidently of intratelluric origin, because some of them occur in the devitrified selvages (Plate 27B).

A few thin tuffaceous and pyroclastic layers are intercalated with the flows south of Ryan Lake, northwest of Lac Gauthier, and north of Lac La Touche. The tuffs are schistose chloritic rocks which contain a few water-worn volcanic pebbles. The pyroclastic rocks consist of irregular, angular, pale green to buff fragments, less than 4 inches long, in an aphanitic, greenish black, chloritic matrix.

Excellent columnar joints can be seen in some of the massive flows (Plate 28A), and a steeply dipping 120 foot-thick flow which crops out three miles north of Lac La Touche shows well-developed two-tier jointing. The lower tier columns are 6 to 15 inches in diameter and the upper tier columns are 10 to 30 inches in diameter (Plate 28B, 28C).

Plagioclase in the volcanic rocks occurs as laths less than 1 mm long. In massive specimens, the laths are randomly oriented and form a network enclosing the other minerals (Plate 29A), suggesting that the original texture was probably intergranular or intersertal. In specimens showing well-developed flow cleavage, original textures cannot be recognized because of mineral growth in the cleavage. Varioles, less than 0.5 mm in diameter, are abundant in the margins of some pillows (Plate 29B).

Composition and Origin

Albite, actinolite, chlorite, clinozoisite, and sphene are abundant minerals in most of the Doublet volcanic rocks. The albite and fine, granular clinozoisite pseudomorph original calcic plagioclase, and both actinolite and chlorite replace pyroxene. Magnetite, ilmenite, leucoxene, quartz, and calcite are commonly present in small amounts. Chlorite and

PLATE 28

Columnar jointing in Doublet metabasalt

A. Columns are vertical, indicating horizontal attitude of the flow at this location. Lac Tantouin in the background.

B. Columnar joints in a steeply dipping flow of Doublet metabasalt. Hammer on one of the columns indicates scale.

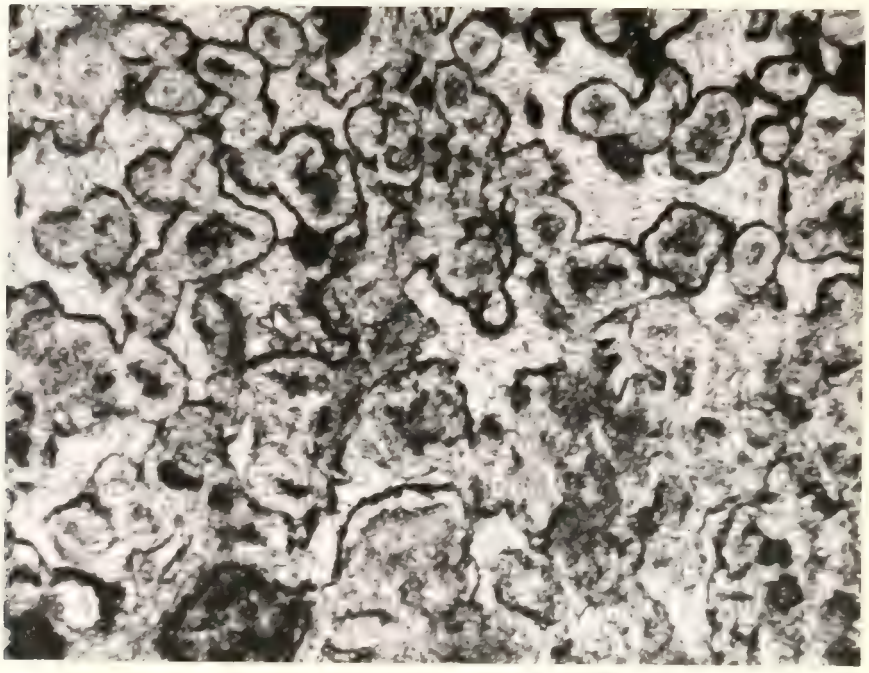
C. Base of a near-vertical flow of Doublet metabasalt. Note cross-sections of columnar joints in the vicinity of the hammer. These are lower tier columns (compare size to A and B).

PLATE 29

Photomicrographs of Doublet metabasalt

A. Thin section showing texture of typical Doublet metabasalt. Acicular laths of plagioclase in a felted mass of clinozoisite, actinolite, chlorite, sphene, and iron oxides. Ordinary light, X30.

B. Thin section showing varioles in a specimen taken from the margin of a pillow. Ordinary light, X30.



carbonate are abundant along shear zones. Approximate modal compositions of two Doublet metabasalt specimens are shown in Table 17, and the chemical composition of a Doublet metabasalt sample from the Willbob Lake area (Frarey, 1954), is shown in Table 18. The sample is similar in composition to Nockold's average tholeiite (Table 9). Soda and titania contents are somewhat low, but this may be atypical for Doublet metabasalts in the Marion Lake area (note high albite and sphene-leucoxene contents, Table 17).

TABLE 17

Modal Composition of Doublet Metabasalt

(Volume percent, based on 600 points for each specimen)

	57-183A	57-225A
Albite	30.3	27.3
Actinolite	47.2	45.5
Chlorite	2.5	6.8
Clinozoisite	15.2	16.3
Sphene, Leucoxene	4.8	3.4
Iron Oxides	- -	0.7

The Doublet metabasalts record a long period of remarkably uniform volcanic activity in the history of the Labrador Trough. Abundant pillow structures indicate subaqueous extrusion for at least 80 percent of the volcanic rocks. Those flows which exhibit columnar jointing, and particularly the one which shows two tier-columnar jointing, may well have erupted sub-aerially. Because individual flows are extensive and of uniform thickness, the magma must have been very fluid.

TABLE 18

Chemical Composition of Doublet Metabasalt*

SiO_2	49.86	Na_2O	1.76
TiO_2	0.53	K_2O	0.43
Al_2O_3	15.07	H_2O +	3.18
Fe_2O_3	2.39	H_2O -	0.34
FeO	9.63	P_2O_5	0.20
MnO	0.37	CO_2	Tr.
MgO	6.34	S	0.15
		Total	99.89

*Composite sample, Willbob area (M. J. Frarey).

Analyst: R. C. Farby, Geol. Surv. Canada

INTRUSIVE ROCKS

Numerous sills of slightly to extensively metamorphosed gabbro, diorite, quartz diorite, peridotite, and related rocks crop out in the western half of the area. Many sills appear to be more than 1000 feet thick, but the presence of sedimentary layers between adjacent sills north of Link and Moss Lakes strongly suggests that unexposed sedimentary septae may also occur within the bodies which have been mapped as thick tabular gabbroic units. All contacts observed in the field are sharp and parallel to the bedding of the intruded sedimentary rocks. The only known discordant intrusion is the gabbro body which transects the Denault formation.

Most of the sills are bounded by sedimentary rocks, but a few coarse-grained gabbroic layers are associated with the Doublet and Keco

metabasalts. Possibly some of the fine-grained, massive greenstones which were mapped as flows are actually intrusive rocks. The sills form prominent ridges, and where drift cover is thin, their continuity and uniform thickness can be readily seen in aerial photographs.

Planar joints parallel to sill contacts (Plate 30A), and compositional layering, also parallel to sill contacts (Plate 30B), indicate emplacement prior to folding. The concordance of the sills further substantiates horizontal emplacement; post-folding intrusions would certainly favor zones of structural weakness. Finally, shear zones within the sills, particularly near the axial planes of folds, indicate that the sills have been folded.

Gabbro

Description

The gabbro is typically a tough, massive, medium-grained rock showing well-developed ophitic texture (Plate 31). It is gray to greenish gray on fresh surfaces and rusty brown to dark gray on weathered surfaces. The major primary minerals, augite and labradorite, are extensively altered to actinolite, chlorite, albite, and clinozoisite. Magnetite-ilmenite intergrowths are pseudomorphed by leucoxene and chlorite. Serpentine, epidote, quartz, and pyrite are the only other important minerals.

The porphyritic gabbro, which crops out north of Lean Lake, west of Marion Lake, and in the northwest corner of the area, contains abundant greenish white glomeroporphyritic clots of altered calcic plagioclase in a dark green, fine-grained matrix of chlorite and actinolite.

PLATE 30

Jointing and compositional layering
in intrusive rocks

A. Well-developed vertical joints in a gabbro sill northwest of Marion Lake. The joints are parallel to the sill contacts.

B. Compositional layering in a gabbro sill east of Lean Lake. Light gray porphyritic gabbro interlayered with fine-grained dark gray gabbro. Layering parallels sill contacts.

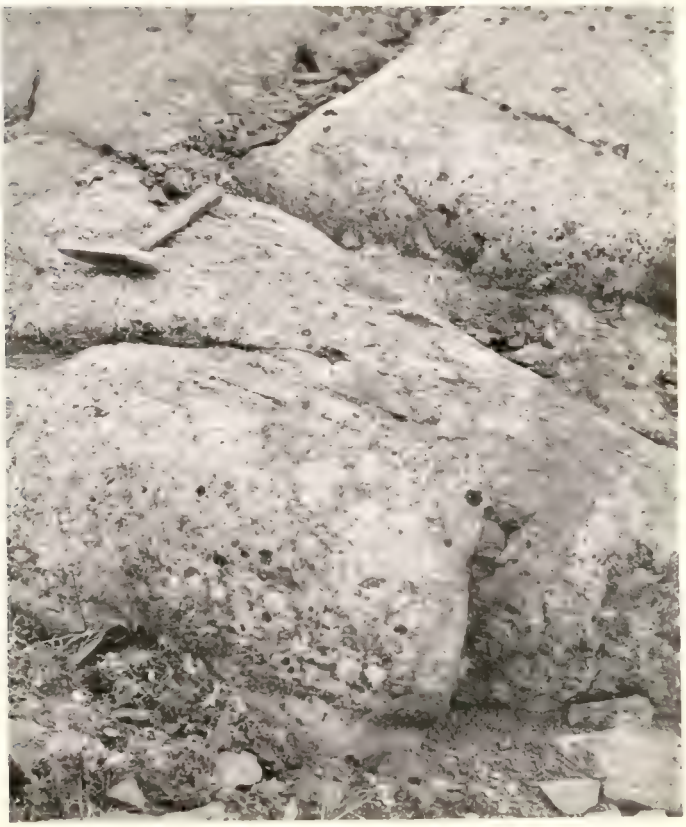
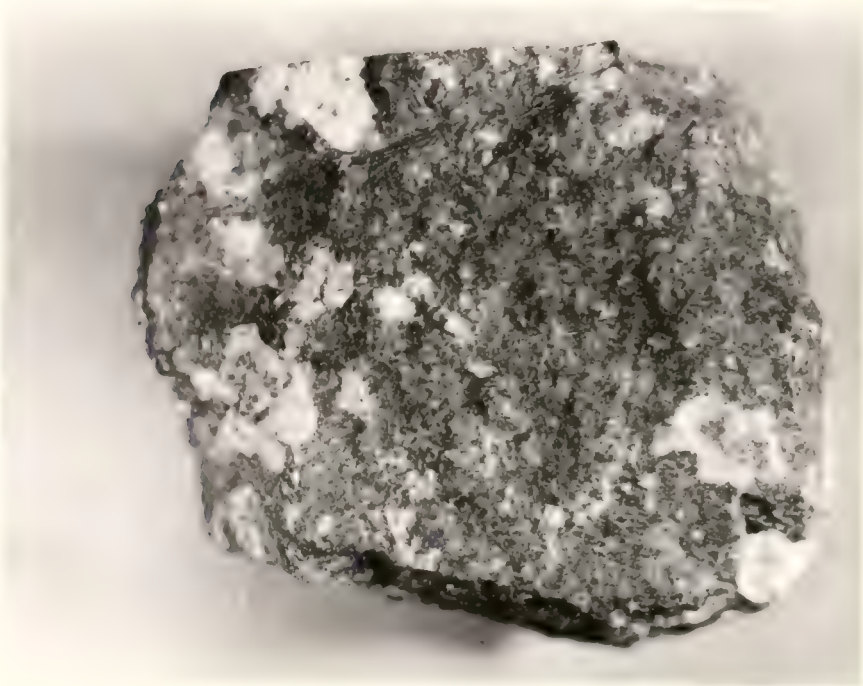


PLATE 31

Gabbro and porphyritic gabbro

A. Thin section showing ophitic texture of typical gabbro. Crossed nicols, X8.

B. Hand specimen of porphyritic gabbro. Note saussuritization of zoned plagioclase phenocrysts. (X1).



The subspherical clots range from 0.05 to 10 cm, and average 2 cm in diameter. In some exposures they compose more than 75 percent of the rock. The porphyritic gabbros probably are intrusive equivalents of the porphyritic Doublet volcanic rocks.

Augite and saussuritized labradorite are abundant in the central parts of the thick sills west of Moss Lake. The augite shows faint pleochroism from neutral to pale brown, and prominent (100) parting can commonly be seen in basal sections. Some large augite crystals exhibit slight zoning, having a (+)2V of about 60° at the core and less than 56° at the rim. The labradorite occurs as narrow euhedral and subhedral laths less than 3 mm long. Broad albite twin lamellae are visible, but are partially masked by clinozoisite.

In most sills, augite has been pseudomorphed by actinolite and chlorite, and labradorite by an almost opaque intergrowth of clinozoisite and albite. Single crystals of actinolite commonly pseudomorph single pyroxene crystals. The actinolite is pleochroic from neutral to pale bluish green. Optic properties include: (-)2V = 75° to 80°; z ∧ c = 15°. The chlorite occurs in fibrous sheaths intimately associated with the actinolite, and also takes the place of magnetite in pseudomorphed magnetite-ilmenite intergrowths. Some chlorite exhibits anomalous blue interference colors, but most has a low, normal birefringence. Both types are pleochroic from pale yellow to yellowish green. Pale greenish brown serpentine, possibly an alteration of enstatite, can be seen in some thin sections. Clear albite rims most of the extensively saussuritized calcic feldspars. Euhedral grains of epidote, as much as 0.5 mm in diameter, occur in some gabbros, but the very fine-grained clinozoisite which clouds the plagioclase grains is more abundant. Clear, anhedral grains of quartz

occupy interspaces between plagioclase laths, and are obviously of primary origin. Intergrowths of magnetite and ilmenite, now replaced by leucoxene and chlorite, also form anhedral patches bounded by plagioclase laths. Granular pyrite, chalcopyrite, and pyrrhotite are disseminated through some of the gabbros. Tiny apatite euhedra occur in the quartz-rich portions of the sills. Thin veinlets containing quartz, chlorite, epidote, calcite, and prehnite are abundant.

Table 19 shows the modal compositions of some representative gabbroic rocks. The chemical composition of a relatively fresh gabbro, calculated from the modal composition, is shown in Table 20.

TABLE 19

Modal Composition of Some Metagabbro Specimens

	57-43A	57-110A	57-118A	57-122N	57-13N
Augite	26.8				
Actinolite	- -	39.5	52.0	34.2	31.4
Labradorite	25.1	- -	- -	- -	- -
Albite Clinozoisite	15.3	38.3	34.8	44.6	49.1
Epidote	3.9	- -	2.3	3.5	- -
Serpentine Chlorite	17.9	7.8	6.4	5.9	15.8
Quartz	5.3	5.2	1.8	6.7	1.6
Pyrite	2.0	1.5	- -	1.5	0.9
Leucoxene	3.7	6.9	2.7	3.6	1.2
Apatite	*	*	- -	*	*

*Less than 0.2 percent.

TABLE 20

Chemical Composition of a Gabbro Specimen,* Calculated from Modal Composition

SiO_2	47.3	MgO	8.8
TiO_2	2.5	CaO	11.0
Al_2O_3	16.4	Na_2O	2.2
Fe_2O_3	2.7	H_2O	1.9
FeO	7.4	Total	100.0

*Specimen 57-43A.

In comparison to the volcanic rocks (Table 9, Table 18) the gabbro has a higher content of lime, magnesia, and alumina, but contains less silica. However, the differences are not great, and may reflect only inaccuracies introduced by the grouping of albite-epidote and serpentine-chlorite. Further, because of the wide range in compositions of the intrusive rocks, one sample can hardly be regarded as representative.

Origin

Evidence has been presented to show that the intrusive rocks were introduced prior to folding (p. 88). It is therefore possible that the gabbros were derived from the same magma as the volcanic rocks. The common occurrence of glomeroporphyritic feldspar clusters in both intrusive and extrusive rocks suggests consanguinity, and as noted above, the chemical compositions are similar.

Diorite and Quartz Diorite

Minor diorite and quartz diorite occur in the upper parts of the gabbro sills within the Marion Lake group and in the upper part of the gabbro sill north of Lac La Touche. These rocks are included with the gabbros because of their limited extent and apparent gradational relationship to the gabbros.

Description

The diorites and quartz diorites commonly show the same texture and structure as the gabbroic rocks, but are lighter in color, ranging from greenish gray to light gray. Albite composes as much as 60 percent of these rocks, but clinozoisite is minor, suggesting alternation of relatively sodic plagioclase. A typical diorite has the following composition: 45% albite, 30% actinolite, 10% chlorite, 5% clinozoisite, 5% quartz, 2% sphene, 2% magnetite, and 1% apatite. A typical quartz diorite contains: 55% albite, 25% quartz, 10% actinolite, 5% clinozoisite, 3% magnetite, 2% apatite. The quartz diorites commonly show micrographic intergrowths of quartz and feldspar, and some contain small amounts of sericitized microcline.

Origin

Occurrence of the diorite and quartz diorite at the tops of gabbro sills strongly suggests differentiation in place. Evidence for separate intrusion is lacking, and similarity of texture, structure, and mineralogy support the differentiation hypothesis.

Serpentinized Peridotite

The ultramafic rocks which crop out on ridges in the vicinity of Lac La Touche form part of a belt which terminates in the Marion Lake area, but extends northwest more than 60 miles. Distribution of the ultramafic rocks indicates concordance with the Doublet rocks and the associated gabbro sills. Compositional layering and planar strike joints are parallel to contacts and several conformable contacts are exposed in the field.

A body of ultramafic rock also crops out a few miles southeast of Lac Rucelai. No contacts are exposed, but its restricted occurrence suggests unconformable relationship to the Doublet metavolcanic rocks.

Description

The ultramafic rocks near Lac La Touche are extremely tough, fine- to medium-grained, and massive. They have greenish black fresh surfaces and orange-brown to dark brown weathered surfaces. Olivine is present in the central part of the sills, but in most places it is completely replaced by serpentine. Tremolite is abundant at the borders of the ultramafic sills, and because of its resistance to weathering, it stands out in relief with respect to the enclosing serpentine.

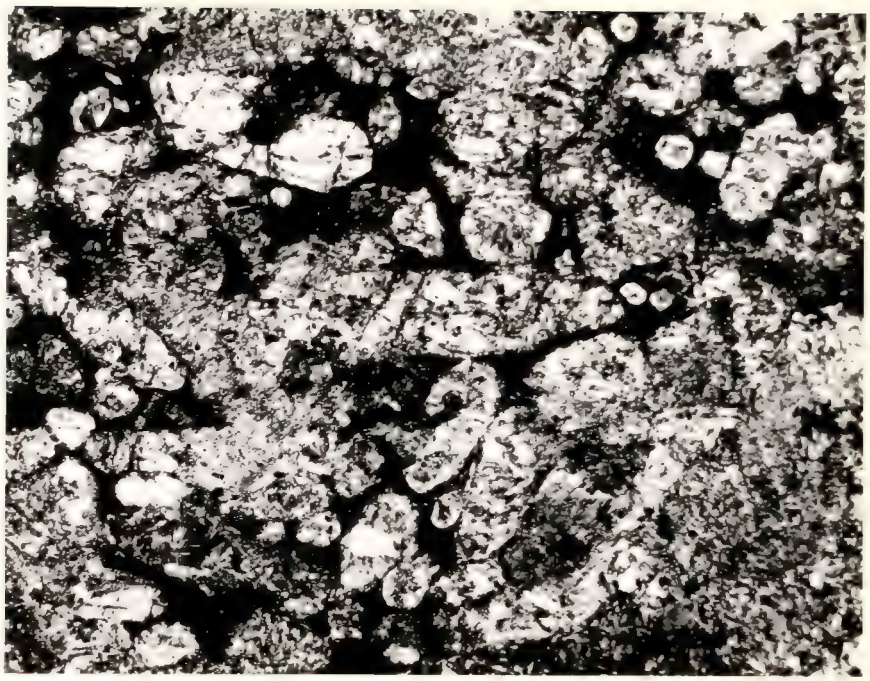
The original texture is well-preserved, and excellent pseudomorphs of olivine are visible in thin section. The olivine, which composed as much as 70 percent of the rocks, is extensively to completely replaced by antigorite, serpophite, and magnetite. The pseudomorphs are 0.25 to 3 mm in diameter, and although most are equant and rounded, some are distinctly tabular (Plate 32A). Remnants of olivine grains can be

PLATE 32

Photomicrographs of ultrabasic rocks

A. Thin section of serpentinized peridotite from the sill east of Lac La Touche. Note alignment of tabular olivine pseudomorphs. Serpophite matrix (black). Crossed nicols, X12.

B. Tremolite-serpentine rock from exposure east of Lac Rucelai. Crossed nicols, X12.



seen in specimens from the central parts of the sills. The antigorite is fibrous and has apparently grown inward from the margins of the olivine grains. Patches of tremolite needles commonly occur in the pseudomorph cores. The magnetite occurs in networks within the olivine pseudomorphs as rounded euhedral grains less than 0.1 mm in diameter. Pale green, near-isotropic serpentine (serpophite) is abundant in most specimens.

Tremolite composes more than 60 percent of specimens from border zones of the sills. It occurs as equant grains, as much as 2 cm in diameter, which are probably pseudomorphs after pyroxene. Most of the tremolite is colorless, but some shows faint green pleochroism, indicating that it is actinolitic. Brown hornblende, a minor constituent of most ultrabasic specimens, shows strong pleochroism from colorless to reddish brown. Marginal alteration to tremolite is common. A fibrous pink to yellowish chlorite which shows anomalous blue interference colors is present in small amounts.

Modal compositions of two ultrabasic rocks are shown in Table 21, and the chemical composition of an ultrabasic sample from the adjoining Griffis Lake area (Fahrig, 1953, p. 46) is shown in Table 22 (A). To better compare the sample with Nockolds' average peridotite (C), the water of metamorphism has been deleted, and the oxides recalculated to total 100 percent (B). The Griffis Lake sample differs from the average peridotite in having a higher alumina content, but the silica, total iron, magnesia, lime, and alkali contents are similar. The mineralogy of the specimens shown in Table 21 does not reflect a high alumina content, suggesting that the Griffis Lake sample may have contained feldspar, augite, or other aluminous minerals.

TABLE 21

Modal Composition of Ultrabasic Specimens

(Volume percent, based on 800 points for each specimen)

	57-254	57-259
Antigorite	59.2	35.1
Serpophite	12.0	9.3
Tremolite-Actinolite	14.3	41.3
Brown Hornblende	8.0	5.9
Chlorite	3.9	1.2
Magnetite	2.6	7.2

The ultrabasic rock southeast of Lac Rucelai consists mostly of tremolite and serpentine. It lacks the pseudomorphic texture so well shown by the Lac La Touche ultrabasic rocks, showing instead a well-developed nematoblastic texture (Plate 32B).

Origin

The ultramafic rocks near Lac La Touche possibly comprise discrete sills, but their distribution with respect to the associated gabbroic rocks suggests that they form the lower portions of differentiated sills. The 1500 foot thick tabular body which borders Lac La Touche must consist of at least 3 sills, because two bands of sedimentary rock crop out within it. Because of thick glacial cover, however, the individual sills could not be mapped.

TABLE 22

Chemical Composition of an Ultrabasic Sample

	A	B	C
SiO_2	38.65	42.0	43.54
TiO_2	.34	0.3	0.81
Al_2O_3	6.59	7.2	3.99
Fe_2O_3	4.24	4.6	2.51
FeO	6.93	7.5	9.84
MnO	.16	.2	0.21
MgO	29.74	32.2	34.02
CaO	3.52	3.8	3.46
Na_2O	.15	.2	0.56
K_2O	.10	.1	0.25
H_2O +	8.35	8.8	0.76
H_2O -	.24		
CO_2	.59		
Cr_2O_3	.08		
NiO	.14		

A. Composite Sample, Griffis Lake area.
Analyst: H. B. Wiik, Helsinki.

B. Sample A, recalculated for comparison with average peridotite.

C. Average of 23 peridotite samples (Nockolds, 1954, p. 1023).

The ultrabasic body southeast of Lac Rucelai may be a sheared and recrystallized facies of the Lac La Touche metaperidotite belt, but its occurrence at the faulted border of the Labrador Trough suggests separate intrusion.

Dike Rocks

A few diabase dikes crop out in the gneissic terrain east of Lac Alonce. The dikes are tabular bodies, 1 to 25 feet thick, which occupy vertical, northwest-trending fractures in the gneiss. Margins of the dikes are aphanitic and melanocratic, but the central parts of the thicker dikes are fine- to medium-grained, mesocratic, and show well-developed diabasic texture. The coarser grained rocks exhibit a gray and green mottling on fresh surfaces. Weathered surfaces are buff to grayish brown. The modal composition of a typical dike rock is shown in Table 23.

TABLE 23

Modal Composition of a Diabase Specimen*

(Volume percent, based on 800 points for each specimen)

Plagioclase	54.7
Titanaugite	10.8
Serpentine	13.3
Chlorite, Actinolite	11.4
Biotite	1.9
Magnetite	7.6
Sulphide Minerals	0.3

*Specimen 4A, one mile east of Lac Alonce.

Plagioclase composes more than 50 percent of the dike rocks. It shows progressive normal zoning from An_{55} to An_{30}, and is commonly clouded by fine-grained sericite. Titanaugite, serpentine, chlorite,

actinolite, biotite, and magnetite are the other abundant constituents. The serpentine pseudomorphs olivine, and, in places, remnants of the olivine can be seen. Chalcopyrite and pyrite are present in small amounts.

GNEISS AND AMPHIBOLITE

The eastern one-third of the map-area is underlain by quartzo-feldspathic gneisses and amphibolites which differ from rocks of the Labrador Trough in structure and grade of metamorphism. Drift cover is thick and bedrock exposures are scarce in much of the low, swampy eastern part of the area, but there are sufficient outcrops to indicate the aerial distribution of the gneiss and amphibolite units (see map in pocket). Most of the amphibolite occupies a belt, 1 to 3 miles wide, which lies along the eastern boundary of the map-area. The gneisses underlie the area between the amphibolite belt and the fault marking the eastern boundary of the Labrador Trough.

Gneiss

Most of the gneisses are medium-grained, well-foliated, leucocratic to mesocratic rocks which commonly exhibit a near-horizontal lineation in the foliation. Alternation of biotite layers a few millimeters thick and quartzo-feldspathic layers less than 1 cm thick is responsible for the foliation; the main linear elements are crenulations and mineral streaks in the foliation. Massive to faintly-foliated biotite diorites, included with the gneiss on the map, occur sporadically as nebulous, lenticular bodies as much as 100 feet long, elongate parallel to the foliation of the enclosing leucocratic rocks. Concordant tabular layers, angular inclusions, lenticular schlieren, and irregular nebulous pods of plagioclase-quartz-hornblende gneiss and plagioclase-hornblende-biotite gneiss are sufficiently abundant in some exposures to lend a migmatic appearance to the rocks (Plate 33A).

PLATE 33

Gneissic outcrops

A. Outcrop showing interlayered leucocratic and melanocratic gneiss.

B. Nebulous pods of melanocratic and mesocratic gneiss enclosed by leucocratic gneiss.

The schlieren are flattened in the foliation and are elongate parallel to fold axes. They have indistinct outlines which interfinger with the enclosing gneiss. Possibly some of them are metamorphic segregations, but because all stages between sharply outlined inclusions and nebulous pods can be seen, and because foliation of the inclusions is commonly discordant with foliation of the enclosing leucocratic gneiss, most schlieren probably originated by successive disruption of the associated tabular bodies. Foliation in many of the nebulous pods passes into the adjoining leucocratic gneiss, indicating incipient replacement by the gneiss (Plate 33B).

The inclusions are commonly enveloped by a relatively mafic-free rim, as much as 3 inches thick, which weathers out in relief. Where such a rim is present, the inclusion border is darker than the core. This relationship suggests that the halos around the inclusions are reaction rims.

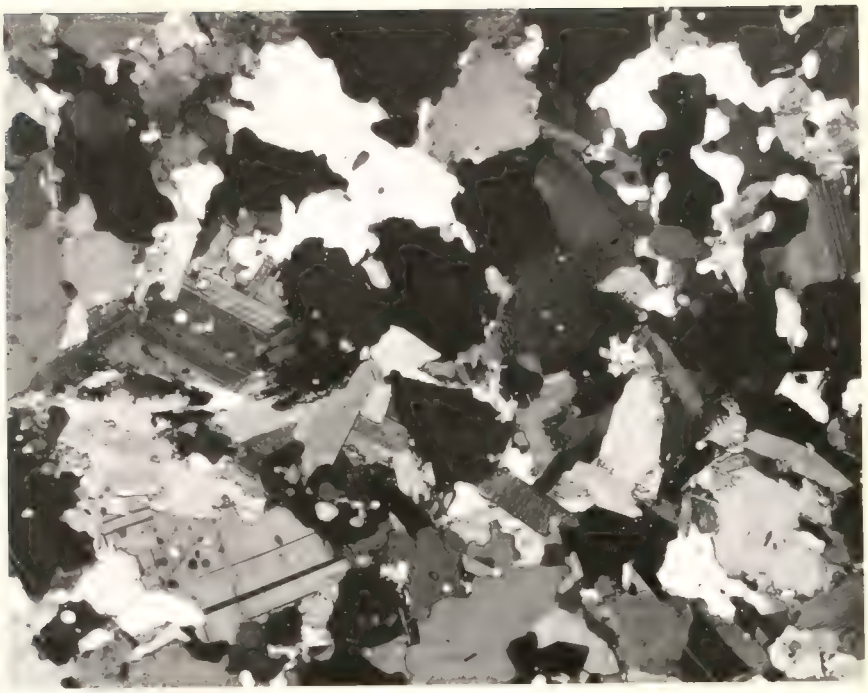
The quartzo-feldspathic gneisses characteristically have a granoblastic to poikiloblastic texture (Plate 34A). Modal compositions of typical specimens are shown in Table 24. Plagioclase, An_{28-32}, occurs as equant, anhedral grains as much as 5 mm in diameter, but most grains are less than 2 mm in diameter. Excellent albite twinning is exhibited by the plagioclase, and pericline twin lamellae are present in some grains. Much of the plagioclase contains abundant euhedral inclusions of apatite, epidote, biotite and zircon, and many grains contain planes of dusty inclusions. Wavy extinction and bent twin lamellae are common features, and in some specimens, the plagioclase is slightly zoned. A few plagioclase grains exhibit incipient saussuritization, but most are clear and fresh.

PLATE 34

Photomicrographs of gneissic rocks

A. Thin section showing granoblastic texture of leucocratic gneiss. The most abundant minerals are quartz, plagioclase, and biotite. Crossed nicols, X12.

B. Thin section of mesocratic gneiss showing typical lepidoblastic to granoblastic texture. Abundant minerals are quartz, plagioclase, hornblende, and epidote. Crossed nicols, X12.



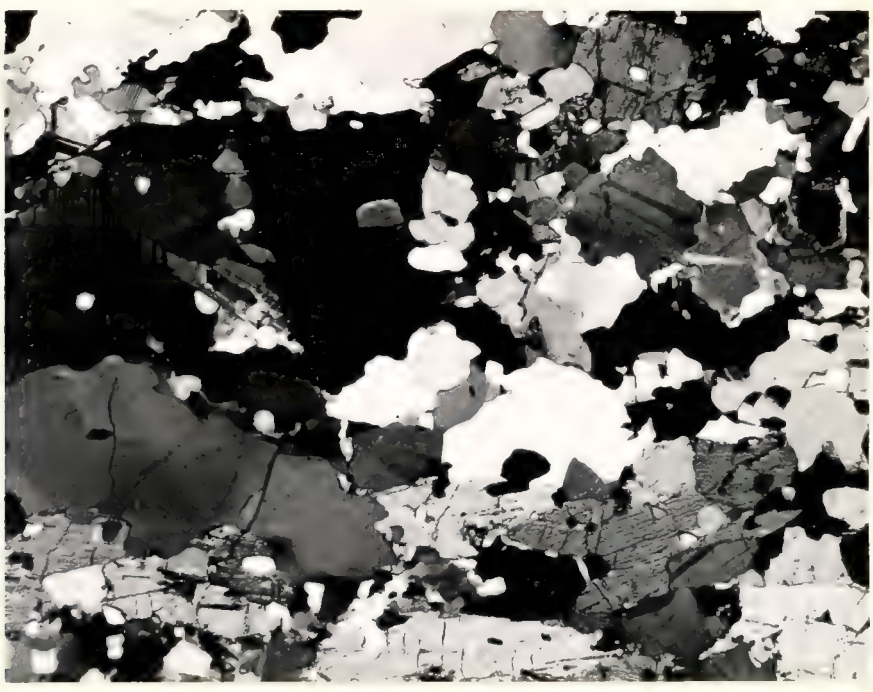


TABLE 24

Modal Composition
of Some Quartzo-Feldspathic Gneisses

(Volume percent, based on 800 points for each specimen)

	DF58-1A	DF58-17A	DF58-58A	DF58-65A	DF58-73A
Quartz	28.2	24.8	34.2	43.2	24.2
Plagioclase	59.7	52.5	38.8	32.5	64.5
Biotite	8.1	13.2	15.3	13.4	5.7
Muscovite	*	*	9.0	1.8	2.1
Chlorite	- -	- -	- -	1.2	- -
Hornblende	- -	3.2	- -	*	- -
Epidote	3.0	3.8	2.3	6.7	2.9
Sphene	0.6	1.5	0.4	*	0.6
Apatite	0.4	1.0	- -	- -	- -
Calcite	- -	- -	0.3	1.2	- -
Iron Oxides	*	- -	*	- -	- -
Pyrite	*	- -	*	- -	- -

*Less than 0.2 percent.

The largest quartz grains are 3 mm in diameter; the average size is 0.5 mm. Undulatory extinction is common, and many of the larger grains are extensively fractured. Most grain boundaries are smooth, but some interfaces between quartz and 001 terminations of twinned plagioclase are sutured. Small ovoid patches of unstrained quartz sieve some of the larger plagioclases. The quartz islands are not in crystallographic continuity.

Biotite occurs as fresh euhedral to subhedral laths with fibrous terminations. Because it does not exhibit bent cleavage or undulatory extinction, and because many laths are disoriented, the biotite probably grew later than the strained quartz and plagioclase with which it occurs. The pleochroic formula for the biotite is: X = pale yellow, Y = greenish yellow, Z = deep olive-brown. There are dark pleochroic halos around tiny inclusions of allanite and zircon in the biotite.

Epidote occurs as euhedral prisms as much as 2 mm long, but smaller anhedral grains are more abundant. Some epidotes enclose rounded cores of orange-colored allanite which are not in crystallographic continuity with the rims. Sphene forms clusters, less than 1 mm in diameter, which in most places are intergrown with segregations of biotite and epidote. Blue-green hornblende, apatite, zircon, and muscovite are minor accessories. In a few outcrops, euhedral garnets, as much as 1 cm in diameter, occur in thin bands parallel to foliation of the gneiss. Large cordierite grains showing ragged margins were observed in one thin section, and euhedral apatite, zircon, magnetite, ilmenite and pyrite are commonly present.

Prominent strain features of the quartz and plagioclase indicate that the gneisses were subjected to extensive cataclasis. Biotite and epidote, however, show no evidence of mechanical deformation, although they mark the foliation. These minerals probably attained their present size after the deformation, growing from seed crystals which were distributed along shear planes during deformation. The rounded boundaries of allanite within many of the euhedral epidotes may well mark the cessation of cataclastic action for this particular mineral.

In Table 25, the chemical composition of a leucocratic gneiss sample is compared with the composition of Nockolds' average muscovite-

biotite tonalite (1954, p. 1015) and Pettijohn's average graywacke (1957, p. 307). The gneiss sample differs from the tonalite in having lower alumina and higher magnesia contents, but it closely resembles the tonalite in the content of other oxides. The sample differs most from the graywackes in having a high silica content and very low iron content. The parent rock of the gneiss could therefore have been a magnesium-rich quartz diorite or an iron-poor and silica-rich graywacke. If metasomatism accompanied metamorphism, other rock types could have formed the gneiss.

TABLE 25

Chemical Composition of a Leucocratic Gneiss Sample

	A	B	C
SiO_2	71.7	70.63	64.7
TiO_2	0.1	0.37	0.5
Al_2O_3	14.1	15.69	14.8
Fe_2O_3	0.3	0.86	1.5
FeO	1.46	1.40	3.9
MnO	0.0	0.04	0.1
MgO	2.0	0.83	2.2
CaO	2.5	2.82	3.1
Na_2O	4.6	4.91	3.1
K_2O	1.0	1.68	1.9
H_2O	0.66	0.62	3.1
P_2O_5	0.0	0.15	0.2
CO_2	0.32	- -	1.3
Total	98.7	100.00	100.8

A. Composite Sample, leucocratic gneiss (58-65A).
Analyst: G. Bender, Geol. Surv. Canada

B. Average of 9 muscovite-biotite tonalites (Nockolds, 1954, p. 1015).

C. Average of 23 graywackes (Pettijohn, 1957, p. 307).

Mesocratic to melanocratic, medium-grained hornblende-rich gneisses are interlayered with the quartzo-feldspathic gneisses in many places. All gradations from sharply-defined concordant layers to nebulous lenticular bodies can be seen in the field. The hornblende-rich rocks exhibit lepidoblastic textures (Plate 34B), and contain abundant hornblende, plagioclase, and quartz (58-2A, Table 26).

TABLE 26

Modal Composition of Some Mafic Rocks
Associated with the Quartzo-Feldspathic Gneisses

(Volume percent based on 800 points for each specimen)

	58-2A	58-20A	58-77A
Quartz	17.4	5.3	8.9
Plagioclase	16.9	58.1	20.5
Hornblende	54.4	0.5	26.8
Biotite	2.9	25.3	38.6
Chlorite	2.8	*	4.3
Clinozoisite	0.4	- -	0.5
Sphene	0.3	*	*
Iron Oxides	- -	6.3	- -
Pyrite	0.8	0.6	0.4
Apatite	*	3.8	*

*Less than 0.2 percent.

58-2A Hornblende-rich gneiss which is interlayered with quartzo-feldspathic gneiss (Sp. 58-1A, Table 24).

58-20A Biotite diorite from a large massive body enclosed by leucocratic gneiss.

58-77A Plagioclase-hornblende-biotite gneiss.

Mesocratic, medium- to coarse-grained, massive and poorly-foliated biotite diorites form pod-like bodies within the quartzo-feldspathic gneisses. Because they are conformable with the enclosing gneiss, exhibit porphyroblastic textures, and show no intrusive relationship to the quartzo-feldspathic gneisses, they are interpreted as local facies of the metamorphic complex. The plagioclase of these rocks has the same range of anorthite content as the plagioclase of the associated leucocratic gneisses. Biotite, the most abundant mineral (58-20A, Table 26) is strongly pleochroic (X = pale yellow, Y = sepia, Z = greenish brown). Small, ragged flakes of biotite form patches at the margins of the large, equant, plagioclase grains. Both plagioclase and biotite commonly exhibit undulatory extinction, but a recrystallized, unstrained generation of biotite is present in some of the biotite diorites. Apatite and magnetite are abundant, and small, sieved, blue-green hornblendes are present in some specimens.

Well-foliated, fine- to medium-grained, plagioclase-hornblende-biotite gneiss composes about 15 percent of the gneissic complex. It has a lepidoblastic texture and contains abundant blue-green hornblende, fresh plagioclase (An_{32-34}), and strongly pleochroic yellowish brown biotite. Tiny metamict zircons are present in the biotite. The modal composition of a typical specimen (58-77A) is shown in Table 26.

Amphibolite

The amphibolites crop out east of Lac Civille and Lac Bonnivet in a north-trending belt approximately 1 mile wide. They are fine- to medium-grained, well-foliated, greenish gray to black rocks (Plate 35) which exhibit an excellent lineation in the foliation. The foliation

PLATE 35

Amphibolite outcrops

Outcrops showing well-developed jointing in the amphibolites. Frost-heaved along one set of joints.

consists of alternate feldspathic and hornblendic layers 0.5 to 2 mm thick, and the lineation is defined by sub-parallel hornblende prisms (Plate 36). All specimens examined in thin section have well-developed lepidoblastic textures. Modal compositions of three typical amphibolites are shown in Table 27.

TABLE 27

Modal Composition of Some Amphibolites

(Volume percent based on 800 points for each specimen)

	58-78A	58-80A	58-82A
Quartz	3.1	- -	- -
Plagioclase	24.2	18.9	16.1
Hornblende	61.5	66.3	70.9
Augite	8.3	7.7	1.8
Clinozoisite	0.8	6.5	10.3
Sphene	2.1	0.4	0.9
Pyrite	*	0.2	*
Apatite	*	*	- -

*Less than 0.2 percent.

Moderately pleochroic hornblende (X = neutral, Y = greenish yellow, Z = green) is the most abundant mineral. It occurs as acicular prisms 0.05 to 1 cm long. The plagioclase (An_{36-42}) is xenoblastic, fresh, and commonly exhibits sharply-defined albite and pericline twinning. Large plagioclase grains show evidence of strain. Prisms of colorless clinozoisite ($2V = 80^{\circ}-95^{\circ}$) are very abundant in some amphibolite specimens, and granular sphene is invariably present. The pyroxene is colorless,

PLATE 36

Photomicrographs of amphibolite

Thin sections showing alignment of hornblende prisms in typical amphibolite specimens. Crossed nicols, X12.

non-pleochroic, and has the prominent (100) parting characteristic of augite. The pyroxene crystals are fresh and poikiloblastic, suggesting growth during metamorphism. Magnetite, pyrite, apatite, and quartz are accessory minerals.

The chemical composition of a representative amphibolite sample and Nockolds' average pyroxene gabbro (1954, p. 1020) are compared in Table 28.

TABLE 28

Chemical Composition of an Amphibolite Sample

	A	B
SiO_2	48.1	50.78
TiO_2	0.6	1.13
Al_2O_3	12.6	15.68
Fe_2O_3	1.9	2.26
FeO	7.15	7.41
MnO	0.1	0.18
MgO	10.7	8.35
CaO	14.7	10.85
Na_2O	1.2	2.14
K_2O	0.2	0.56
H_2O	1.22	0.48
P_2O_5	0.0	0.18
CO_2	0.06	- -
Total	98.5	100.00

A. Composite amphibolite sample (58-82A).
Analyst: G. Bender, Geol. Surv. Canada

B. Average of 38 pyroxene gabbros (Nockolds, 1954, p. 1020).

The amphibolite sample has higher lime and magnesia contents, and lower silica, soda, titania, and alumina contents, but a slight difference in composition of the pyroxene could account for the differences. None of the common sedimentary rocks compositionally resemble the amphibolite. The amphibolites were therefore probably derived from sills or flows of gabbroic composition.

STRUCTURAL GEOLOGY

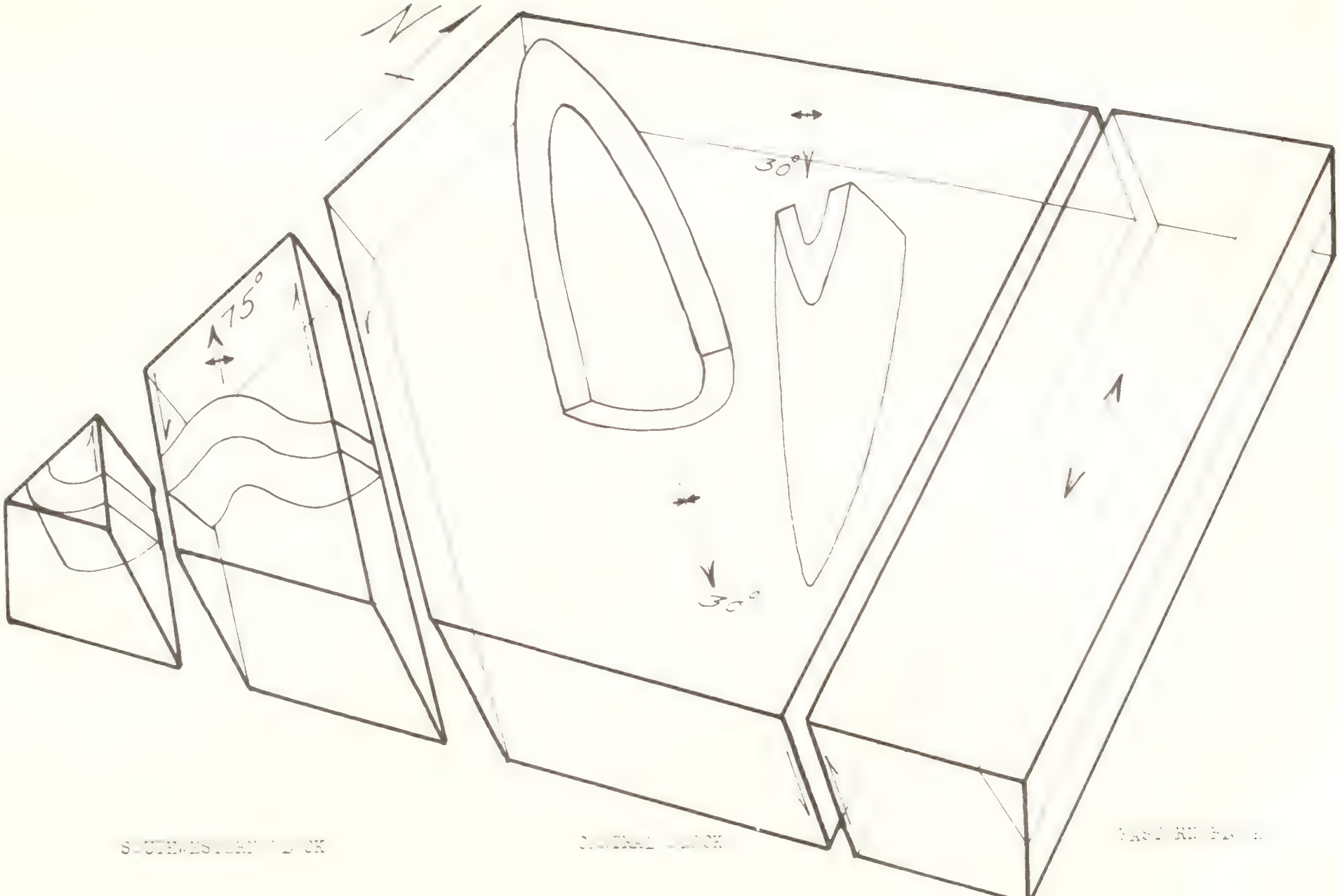
The western two-thirds of the Marion Lake area is underlain by rocks of the Labrador Trough. The eastern one-third is underlain by gneiss and amphibolite. Folds and major faults in the Trough trend northwest and are truncated by gneisses. Because of this structural discordance, a major fault, along which the eastern block moved up, must separate the Trough rocks from the gneisses and amphibolites. Another prominent fault separates the Knob Lake group, in which fold axes plunge steeply northwest, from the Murdock and Doublet groups, in which fold axes plunge gently southeast. The Marion Lake area can, therefore, be divided into three structurally distinct units. For convenience, the three units will be referred to as the Southwestern Block, the Central Block, and the Eastern Block (Fig. 10). Two geologic cross-sections (pocket at back) show the structural interpretation.

Southwestern Block

The metasedimentary formations of the Knob Lake group are structurally characterized by fold axes which plunge steeply northwest. The outcrop pattern outlines a large syncline and anticline, and small-scale structures which are congruent with these large folds have been observed in many outcrops.

Fig. 10

Transparent block diagram showing the three structural units in the Marion Lake area. The basic structural elements are shown schematically.


N
30°
75°
35°

Folds

Folds in the Knob Lake group are accentuated by the metagabbro sills northwest of Marion Lake and by both sills and flows in the vicinity of Mountain and Keco Lakes. Noticeable, however, is the disparity between outcrop patterns of the Marion Lake formation and the overlying formations (Keco, Denault, Wishart, and Menihek). On the west shore of Marion Lake, the Keco and Marion Lake formations are only a few hundred feet apart; one mile west of the shore, they are separated by a distance of more than two miles. Metagabbro occupies the wedge between them. Undetected faults may be present, but laccolithic intrusion of the gabbro affords a more likely explanation. Configuration of the gabbro and its conformable relationship to the enclosing formations strongly suggest intrusion accompanied by arching of overlying formations with respect to the Marion Lake formation.

Drag-folds are abundant in the Menihek formation, and many of them are congruent with the large-scale folds. However, drag-folds have also been developed along shear zones, and care must be taken to relate them to the correct large-scale structures.

Faults

All faults which were mapped in the field can be recognized as strong lineaments on aerial photographs. In the field, straight shallow valleys commonly mark the fault zones. Rock in the fault zones is highly sheared or fractured, but crops out in few places because drift cover is thick in the fault-controlled valleys. Prominent lineaments for which no field evidence of faulting was obtained are shown on the map by means of a separate symbol.

The fault separating the Knob Lake and Murdock group has been mapped for more than 50 miles to the northwest (Frarey, 1952), but has not been investigated southeast of the Marion Lake area. Near-vertical drag-folds and near-horizontal slickensides indicate transcurrent movement with right hand offset. Another transcurrent fault with right hand offset passes through Kecc Lake, and parallels the above mentioned fault.

Cleavage and Lineation

Flow and fracture cleavages are well-developed in the Attikamagen and Menihek rocks and fracture cleavage is prominent in the siltstones of the Marion Lake formation. The flow cleavage in most places trends northwest and is steeply dipping. Because the axial planes of large folds have the same attitude, this cleavage is presumed to be related to folding (De Sitter, 1956, p. 99), and is therefore an axial plane cleavage. Flow cleavage in highly folded areas is gradational with fracture cleavage in less deformed rocks, showing that it developed from fracture cleavage, but in many places a second fracture cleavage transects the flow cleavage. Too few instances of this later fracture cleavage were observed to determine whether it is also related to folding. In many places, cleavage and bedding intersect to form a lineation which plunges steeply northwest. Because this lineation parallels axes of folds, it is a b-lineation (Sander, 1926, p. 328). Cleavage is absent in the Marion Lake sandstones, Denault formation, and gabbros, but fracturing sub-parallel to axial planes is common.

Joints

Planar joint sets are well-developed in the sills. Joints parallel to the sill contacts are most prominent, but several sets perpendicular to the contacts are also present. Because of this relationship, the joints are an aid in structural mapping.

Cross joints (ac joints) are present in parts of the Attikamagen, and are particularly well-developed in the slates which crop out on the east shore of the southwest arm of Marion Lake. Joints are also conspicuous in much of the Menihek formation, and commonly there are two sets sub-perpendicular to the bedding. In places, these joint sets intersect at about 60 degrees, forming large rhombic slabs which, in the steeply-dipping beds, tend to upthrust as a result of frost action.

Joints are abundant in the Denault formation. Many joint surfaces are sub-perpendicular to bedding and show no apparent relation to fold configurations. Schmidt (1957) measured more than 9000 joint surfaces in folded carbonate rocks of the Central Foothills of Alberta, and found that the joints intersected bedding at high angles, regardless of position in folds. Because of this apparent lack of structural control, he concluded that the joints developed while the strata were essentially flat lying. A statistical study of joints in the Denault was not made, but sufficient joint-bedding relationships were observed to strongly suggest the situation described by Schmidt.

Central Block

Large-scale folds, longitudinal faults, and shear zones are abundant in the Doublet group. Most of the Murdock group is highly schistose, and no large-scale folds have been recognized within it.

Folds

Axes of the numerous folds in the Doublet group trend northwest to north-northwest, converging towards the southeastern part of the Central Block. Axial planes dip steeply northeast to east, and the limbs of some folds are locally overturned towards the southwest. Most fold axes plunge gently southeast, but the axes of the folds between Lac Rucelai and Lac Godefroy are horizontal. Most of the plunging folds can be recognized and traced on aerial photographs, but the horizontal folds are recognized only by reversals in dip where the upper sides of flows can be identified. Flow cleavage or schistosity is well-developed in the Doublet metavolcanic rocks. The cleavage is sub-parallel to axial planes of folds and commonly contains elliptical chlorite smears which are elongate in "a". The relationship of cleavage, lineations and folds is identical to that described by Cloos (1950, p. 8) for South Mountain Anticlinorium in Maryland.

Faults

Major faults within the Doublet group are marked by shear or fracture zones as much as 300 feet wide, commonly occupy straight, shallow valleys, and show as lineaments on aerial photographs. They converge, as do the folds, toward the southeastern part of the Central Block. Fracture cleavage, which is prominent in most fault zones, dips northeast at lesser inclination than the flow cleavage. Intersection of flow and fracture cleavages commonly forms a b-lineation which parallels fold axes. In many fault zones, vertical transport is indicated by slickensides on the frac-cleavage surfaces. Chlorite is abundant in the shear zones, and in some

its growth parallel to shearing has resulted in development of a secondary schistosity which masks the regional schistosity.

Eastern Block

Foliation

A well-developed foliation is the most prominent structure in the Eastern Block. The foliation is steeply inclined in most places, and has a remarkably uniform northward trend within the outcrop belt. The strike of the foliation parallels the inferred contact of the Eastern and Central Blocks. Consistent reversals in dip suggest that the foliation is isoclinally folded.

Foliation in the gneisses consists of alternate dark- and light-colored layers less than 1 cm thick. Where subparallel flakes of biotite are abundant in the dark layers, the gneisses tend to fracture in flaggy slabs, but in most outcrops, the gneiss breaks along irregular surfaces unrelated to the gneissosity. Lenticular mineral aggregates, mafic schlieren, and tabular bodies of melanocratic gneiss which are commonly interlayered with the leucocratic gneisses accentuate the foliation in many exposures.

Foliation in the amphibolites consists of interleaved hornblende-rich and plagioclase-rich layers. The layers, which are less than 2 mm thick, probably formed by metamorphic segregation. Some of the amphibolites exhibit a crude fissility parallel to foliation.

Lineation

Linear elements in the gneisses and amphibolites include crenulations on foliation surfaces, fold axes, acicular hornblende prisms,

lenticular pods of mafic minerals, schlieren, and the intersection of foliation and fracture cleavage. All lineations lie in the north trending foliation and are horizontal or plunge gently north. The concordance of lineations in the gneisses and amphibolites suggests a common structural history for the two rock types. Coincidence of fold axes with the regional lineation of the Eastern Block indicates that the linear elements mark a b-lineation.

Joints

Two joint sets are present in the Eastern Block. One set trends northwest and the other trends west; both are vertical to steeply dipping. Because the east-west joints are regionally developed within the gneisses and amphibolites and are perpendicular to the b-lineation, they probably are ac-joints, formed at the same time as the foliation and lineation. The northwest joint set is best developed in the gneiss east of Lac Alonce, where diabase dikes have been intruded along some of the joints. Contraction fractures normal to the dike walls are developed here and there, but the dikes are nowhere offset by younger fractures. A northeast joint set is present in outcrops of amphibolite 2 miles east of Lac Gouffier. Perhaps the northeast and northwest joint sets form a conjugate system.

Relationship of Structural Elements

Structures within the Trough rocks appear to be related in origin. The regional schistosity is an axial plane cleavage which developed during folding. Because the fracture cleavage in fault zones offsets the schistosity or flow cleavage, faulting must have taken place

after folding. However, the faults are sub-parallel to axial planes of folds, and therefore common origin is strongly suggested. Probably forces directed from the northeast caused folding, local overturning to the southwest, and development of flow cleavage, and during late stages of the deformation the faults formed as a direct result of the same southwest-directed force. Reverse faulting along the northeast-dipping faults then relieved the stresses imposed on the Trough rocks, and fracturing in the fault zones obscured the fold-developed flow cleavage. Some transcurrent movement occurred in some of the fault zones, as shown by the right hand offset on the fault in the southwest corner of the map-area. Major transcurrent movement along the fault between the Knob Lake and Murdock groups may have caused juxtaposition of the blocks showing divergent plunge. Faulting at the east margin of the Trough occurred after cessation of Trough deformation.

METAMORPHISM

Regional Metamorphism

The metamorphosed sedimentary, volcanic, and intrusive rocks in the western portion of the map-area belong to the greenschist facies. The gneisses and amphibolites in the eastern portion of the map-area belong to the amphibolite facies of metamorphism. There is no indication of progressive increase in grade of metamorphism from west to east. The greenschist rocks to the west form part of the Labrador Trough, which is believed to be of Proterozoic age. The gneisses and amphibolites may be either Archean basement rocks which have been elevated by faulting, or high grade Trough rocks which have been faulted against low grade Trough with consequent removal of the zone showing increasing grade of metamorphism.

Greenschist Facies Rocks

Both muscovite-chlorite and biotite-chlorite associations occur in the Knob Lake, Murdock, and Doublet groups, but because the mineral assemblages are not zonally distributed, sub-facies (Turner, 1948) cannot be delineated within the Marion Lake area.

Original textures of the metasedimentary rocks are well-preserved only in competent formations. The quartzites, siltstones, and calcareous rocks show little more than slight recrystallization, but most of the pelitic rocks are extensively recrystallized, and show typical greenschist mineralogy. Quartz-albite-chlorite-muscovite and quartz-albite-biotite-chlorite assemblages are common in the slates and phyllites, but clastic oligoclase grains are preserved in some of the sandstones. Perhaps deformation of the less competent pelitic rocks was responsible for their more extensive alteration.

Albite-actinolite-clinozoisite-chlorite-sphene assemblages in the metabasalts and metagabbros have largely taken the place of calcic plagioclase-pyroxene-magnetite-ilmenite assemblages. However, the central parts of some thick gabbro sills are relatively unaltered, showing that equilibrium was not fully attained during metamorphism. Perhaps lack of penetrative deformation in the competent gabbro sills was responsible for preservation of original pyroxene and plagioclase, but because the sill margins are invariably altered, access of water was probably a more important factor.

Amphibolite Facies Rocks

The gneisses and amphibolites consist essentially of plagioclase (An_{28-42}), blue-green hornblende, biotite, quartz, clinozoisite, and sphene,

This mineralogy is typical of the amphibolite facies. Some amphibolites contain as much as 8 percent augite, suggesting a somewhat higher degree of metamorphism, but because abundant hornblende, clinozoisite, and sphene are in equilibrium with the pyroxene, these rocks also belong to the amphibolite facies.

Contact Metamorphism

Intrusion of the sills has had little apparent effect on the country rock, but this may be due to masking of contact metamorphism by the regional metamorphism. All contacts observed in the field are sharp and mineralogy of the intruded rock at the contacts is identical to that of the intruded rock remote from the contacts. A fine-grained, weathering-resistant zone less than 2 inches thick which commonly occurs in country rock at the contacts probably formed as a result of local increase of temperature during intrusion.

Several outcrops of siltstone in the Marion Lake formation show a peculiar mottling consisting of scattered patches rich in chlorite and muscovite. The patches are less than 0.5 cm in diameter and commonly coalesce to form irregular networks parallel to bedding. The spaces between the mottles are invariably bleached, and total content of chlorite and muscovite in mottled beds does not differ significantly from the chlorite-muscovite content of mottle-free beds. Occurrence of the mottled beds near gabbro sills suggests that the mottling is a contact phenomenon.

The tremolite zones at the margins of ultrabasic sills probably formed by contact metamorphism. No analyses are available to demonstrate chemical differences, but the abundance of tremolite suggests a higher content of silica than the serpentine-rich portions of the sills. Perhaps

silica was derived from the sedimentary rocks which are common at contacts where the tremolite zones occur.

Fault Zone Metamorphism

Regional cleavage commonly is obscured by fracture cleavage in the fault zones, and in some places a secondary schistosity has developed parallel to the shearing in the fault zones. Chlorite and carbonate minerals, particularly ankerite, are abundant in the fault zones. The carbonatized rock east of Lac Charny and Lac Buzançois mark zones of extensive shear which are probably faults. Some specimens contain as much as 30 percent ankerite.

Age of Metamorphism

Metamorphism of the Trough rock obviously took place after emplacement of the gabbro sills, and because the high grade gneisses and amphibolites east of the Trough boundary show no signs of retrogressive metamorphism, the Trough metamorphism must have occurred prior to the faulting at the east margin. Because deformation and metamorphism are commonly synchronous, folding of the Trough rocks may well have coincided with the greenschist metamorphism. Development of metamorphic minerals in the regional cleavage, which in turn is related to the regional folding, lends strong support to this possibility.

Metamorphism of the gneisses and amphibolites is closely related to their origin. If these rocks are part of the basement complex, their present mineralogy may have developed prior to deposition of Trough rocks, but it could also have developed either during Trough metamorphism or during the period of marginal upthrust. If the gneisses and amphibolites

are high grade Trough rocks, the age of metamorphism throughout the rocks in the Marion Lake area probably was coincident.

GEOLOGIC HISTORY

Downwarping of the Archean basement initiated sedimentation in the linear belt now known as the Labrador Trough. Sediments derived from bordering highlands formed the basal Seward conglomerates and arkoses which are not exposed in the map-area, but which do crop out along the west side of the Trough. Continued erosion reduced elevation of the borderlands, and deposition of the finer grained Attikamagen clastics ensued. Variations in oxidation potential and organic activity resulted in local deposition of black shales. Uplift along the west side of the Trough during Marion Lake times resulted in renewed deposition of arkoses and conglomerates. Some rock fragments in these sediments were derived from the underlying Seward and Attikamagen. Slight fluctuations in sea level caused shifts of the strand line across a low-lying plain, allowing deposition in shallow marine, littoral, fluviatile, and perhaps aeolian, environments. A period of volcanic activity followed, resulting in extrusion of the Keco flows, but probably no major changes in environment occurred, because further shelf sedimentation followed. The extensive carbonate deposition of Denault times took place in a shallow water to intertidal environment. Algae modified the environment, played a major role in sediment fixation,and may have precipitated much of the dolomite. Some erosion of the Denault occurred before influx of the Wishart clastics. The Wishart sandstone beds must represent periods of time sufficiently long for attainment of the remarkable textural and compositional maturity which they exhibit. Iron sedimentation started in Wishart times, perhaps

as a result of change to tropical climate, or perhaps because of contemporaneous vulcanism. Menihek deposition probably occurred in somewhat deeper water in a reducing environment.

The Murdock rocks represent a period of volcanic activity, dominantly pyroclastic. Sediments of the Doublet were deposited in shallow waters, and although the environment may not have differed greatly from that of Knob Lake times, the Doublet sands were not subjected to extensive abrasion and sorting as were the Marion Lake and Wishart sands. Volcanic activity ensued, accompanied by intrusion of the abundant basic sills of the map-area.

Force directed from the northeast resulted in folding about northwest-trending axes in the Labrador Trough. Cleavage and reverse strike-faults formed during the same period, accompanied by metamorphism to the greenschist facies. The east side of the Trough was later faulted, resulting in uplift of the gneisses and amphibolites which may be either Archean basement rocks or high grade Trough rocks.

ECONOMIC GEOLOGY

Numerous gossans mark the locations of sulphide deposits in the map-area. Some gossans are more than 2000 feet long, but the sulphide bodies from which they were derived are commonly less than 100 feet long. The transported gossans extend downhill from their source; many of the gossans form stream beds. Pyrite is the most abundant sulphide mineral, but chalcopyrite and pyrrhotite are commonly present. The sulphide minerals are disseminated or occupy fractures in the host rocks, which in almost all of the deposits, are slates of the Menihek formation of Doublet group. Shear zones and the contacts between slates and mafic intrusions

are the most common sites of mineralization. Because the slates normally contain finely divided sulphide minerals, it is possible that the sulphide deposits originated by thermal metamorphism (caused by mafic intrusions) and dynamic metamorphism (within shear zones) of the slates.

Veinlets of cross-fibre asbestos occur in the serpentinite bodies near Lac La Touche and in the tremolite-serpentine body southeast of Lac Rucelai. Most of the material is hard and non-flexible; asbestos of commercial grade was not observed.

Quartz veins as much as 10 feet thick crop out west of Lac Gauthier and along the chain of lakes northeast of Moss Lake. Traces of disseminated pyrite and chalcopyrite are present in some of these veins.

APPENDIX A

Calculation of Grain Size Distribution, in Volume Percent Frequency, from Thin Section Data

The grain size distribution of a poorly indurated sedimentary rock is most easily determined by sieve analysis of a representative sample. However, for well-indurated samples, this method cannot be used unless the constituent grains can in some way be separated. For carbonate-cemented quartz sandstones, acid treatment is sufficient for disaggregation, but for rocks in which the cement is identical in composition to the framework particles, no practical method of separation has yet been devised. Alternatively, a selected size parameter can be statistically measured in thin section. Commonly the apparent long axes of grains are measured and grouped in size classes (Krumbein and Pettijohn, 1938, p. 128). The data so obtained, however are expressed as number frequencies, and are not directly comparable with the weight frequency data obtained by sieve analyses.

The relationship between thin section data and sieve data has been considered by many workers, and mathematical correlation factors have been proposed (Krumbein, 1935; Greenman, 1951). A graphical method of translating thin section distributions to sieve distributions described by Friedman (1958) is more convenient, and possibly more accurate, than the mathematical methods.

Because most sandstones in the Marion Lake area are not amenable to sieve analysis, frequency data were obtained for thin sections of the sandstones, and the data converted to sieve equivalents by use of Friedman's graph (p. 414). Frequency distributions were determined with the aid of a microprojector (Krumbein and Pettijohn, 1938, p. 374) instead

of the point count method suggested by Friedman. The microprojector method, which is more convenient and quicker than the point count method, is described below:

1. The microprojector is adjusted to give a suitable image of the thin section on a sheet of white paper. The scale factor is determined by placing a transparent millimeter rule over the thin section.

2. The long axes of all grains in several random fields are measured and grouped according to phi size classes. As each grain is counted, the grain image is stroked off on the field of projection. The numbers in each class are conveniently recorded with the aid of a tally counter. The data are converted to number frequency percent.

3. The frequency of each class is weighted by multiplying by the square of the class mid-point. This step converts number frequency to apparent area frequency. On the assumption that the thin section represents a random cut through the sampled rock, the apparent area frequency is directly proportional to the apparent volume frequency ($V = A\int dx$).

4. The apparent volume frequencies are converted to percent values and then summed to give apparent cumulative percent frequencies for each class.

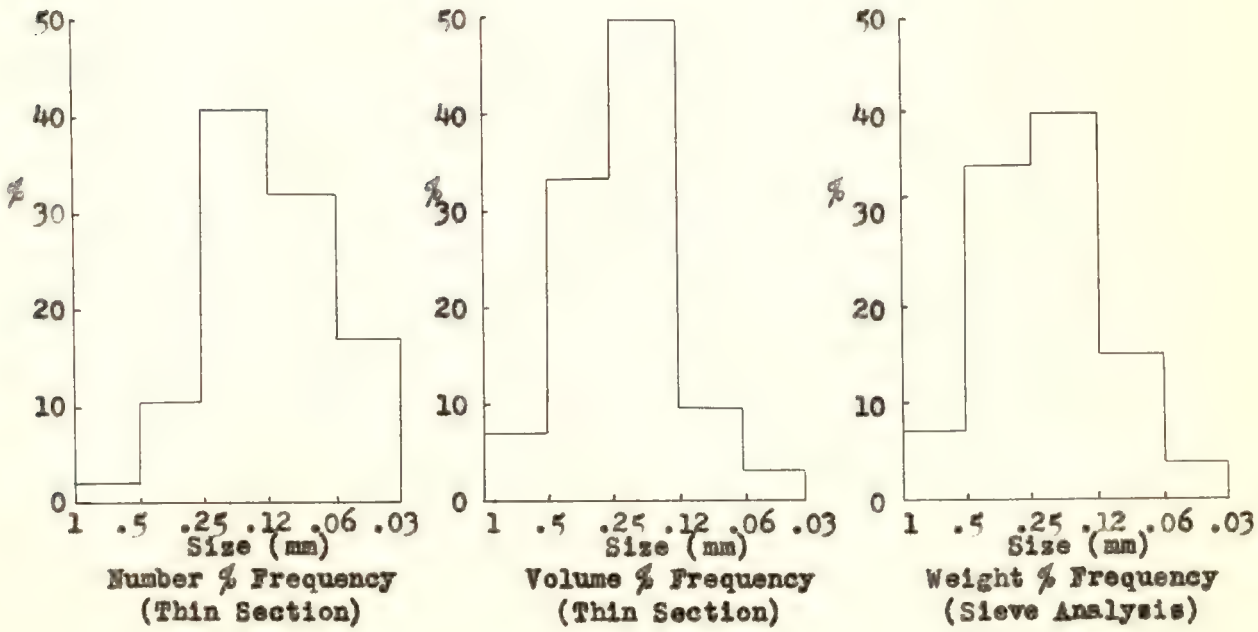
5. The apparent cumulative values are plotted on the graph given by Friedman (1958, p. 414), and the "true" cumulative values are read on the appropriate scale. The "true" volume frequencies are obtained from this set of figures, and plotted in histograms.

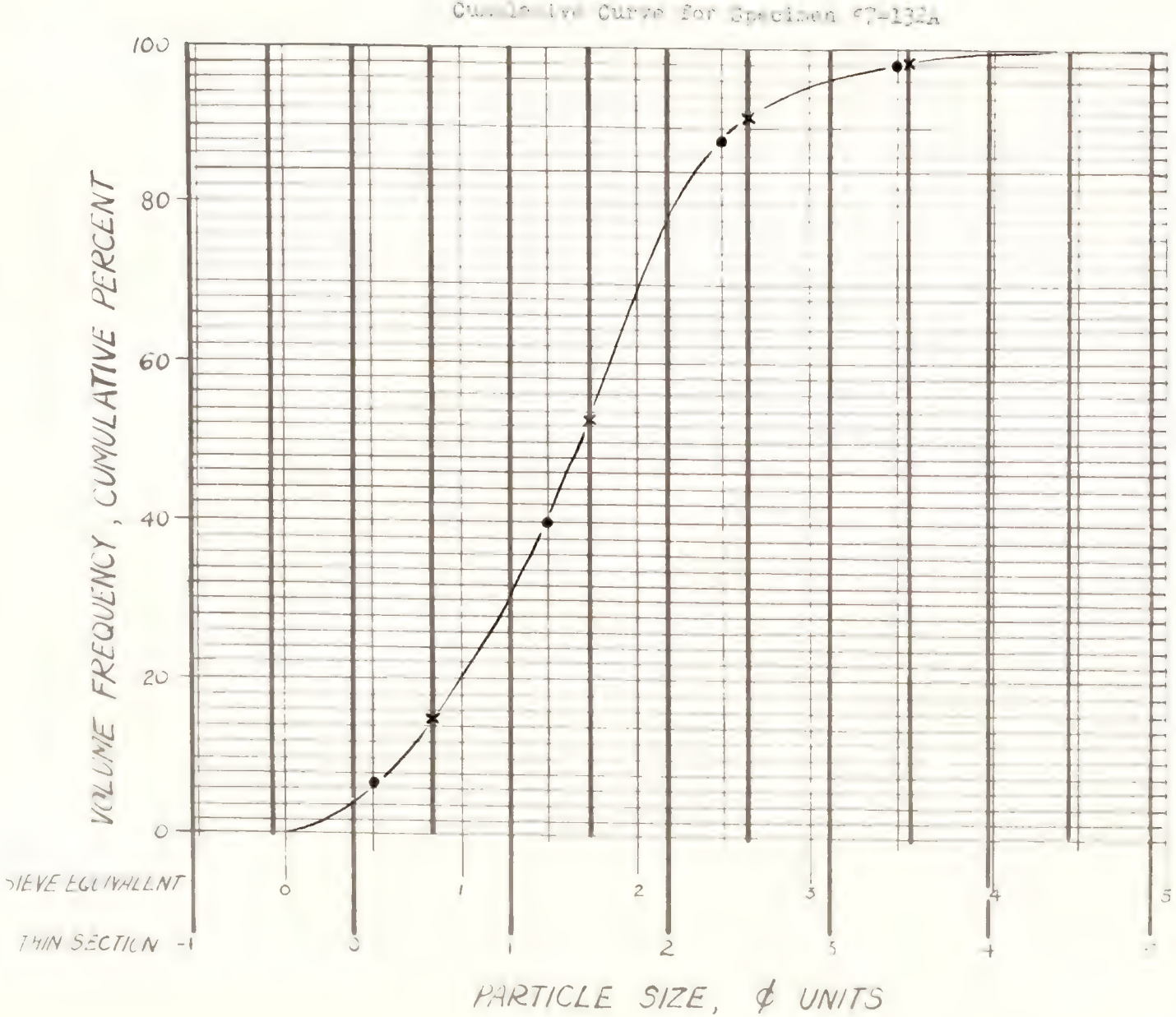
An example of the calculations for a representative Marion Lake sandstone, Specimen 57-132A, is shown on the following page. Because the sandstone from which the thin section specimen was collected is friable, a sample suitable for sieve analysis could also be obtained. The weight distribution, determined by sieving 200 grams of the sample, is shown for comparison with the thin section data.

Example of the Calculation of Grain Size Distribution, in Volume Percent Frequency, from Thin Section Data (Specimen 57-132A)

Class Limits	0	0	1	2	3	4	5
	mm	1	.5	.25	.125	.06	.03
Class Mid Points (d)		.75	.375	.187	.093	.045	
d^2		.56	.141	.035	.009	.002	
Number % Frequency		1	10	40	31	18	
Volume Frequency $_u$		.56	1.41	1.40	.259	.036	
Volume % Frequency $_u$		15.3	38.6	38.3	7.1	0.7	
Volume % Cum. Freq. $_u$		15.3	53.9	92.2	99.3	100	
Volume % Cum. Freq. $_c$		7	40	89	99	100	
Volume % Frequency $_c$		7	33	49	10	1	
Weight % Frequency (Sieve Data)		7.5	35	40	15	2.5	

u = uncorrected data
c = corrected data
cum. = cumulative




Cumulative Curve for Specimen 67-132A
100
80
60
40
20
0
VOLUME FREQUENCY, CUMULATIVE PERCENT
SIEVE EQUIVALENT
0
1
2
3
4
5
THIN SECTION -1
0
1
2
3
4
PARTICLE SIZE, ϕ UNITS

APPENDIX B

Sandstone Size Data

Sample No.		Class Limits (mm) 2	1	0.5	0.25	0.125	0.06	0.03	
57-51A	A		11	32	25	22	10		
	B		19	65	13	100			
57-130A	A		8	16	26	38	12		
	B		21	60	15	4	2		
57-132A	A		1	10	40	31	14	4	
	B		7	33	49	10	1		
57-133A	A			4	28	52	10	6	
	B		2	16	52	28	2		
58-6-4	A		13	17	28	32	10		
	B		24	67	7	2			
58-6-16	A		21	23	33	15	8		
	B		30	59	9	2			
57-116A	A	2	6	17	21	32	12	10	
	B	16	41	30	10	2	1		
57-216A	A	1	3	8	19	27	24	18	
	B	15	41	24	13	5	2		

A. Number frequency, thin section.

B. Corresponding corrected volume frequency.

APPENDIX B

Conglomerate Size Data

Sample No.		Class Limits (mm) 32	16	8	4	2	1	0.5	0.25	0.125	0.06
57-136C*	A		1	9	19	71					
	B						2	8	18	33	39
	C		3	15	14	11	9	9	6	5	1
57-153A**	A			2	21	77					
	B						1	6	22	40	31
	C		4	9	20	14	13	11	8	2	

A. Number frequency, polished surface.
B. Number frequency, thin section.
C. Corresponding corrected volume frequency.

* Area of particles > 2 mm:area of particles 2-0.06 mm:area of matrix = 33:40:27
** Area of particles > 2 mm:area of particles 2-0.06 mm:area of matrix = 34:48:18

APPENDIX C

Geologic Sections in the Denault Formation

The five sections described on the following pages were measured with a 200 foot tape along lines established by Brunton compass. The tape was moved forward in 200 foot increments, and the contacts between units were recorded to the nearest 5 foot mark. The percent exposure, measured thickness, and true thickness of each unit are tabulated in columns adjacent to the lithologic descriptions. The section locations are shown in Fig. 6.

SECTION 1

Unit	Measured Distance	True Thickness	Percent Exposure	Description
19	2355-2740	365	100	Siliceous dolomite: gray, orange-buff weathered; well-laminated, as seen on ribbed weathered surface; fractures smoothly to give a massive-appearing fresh surface; very finely crystalline; laminations planar and persistent laterally; sparse chert nodules which are ovoid in section, elongate parallel to bedding.
18	2300-2355	50	100	Dolomite: gray, buff-gray weathered; finely crystalline; undulatory stromatolites and minor quartz veinlets.
17	2265-2300	30	40	Algal unit; light-gray and buff, carbonaceous dolomite; 4-7" columnar stromatolites, differentially weathered; 1-2 cm pisolites at base; thin veinlets of quartz are abundant.
16	2195-2265 2115-2185	65 60	40 70	Dolomite: gray, buff weathered; medium-grained crystalline; undulatory stromatolites; quartz interlayers near top. Bedding 90;155 N.
	1915-2115	185	70	Dolomite: dark gray, buff-gray weathered; finely crystalline; irregular, undulatory stromatolites in places; veinlets and lenses of quartz abundant; a few chert nodules, and some black chert beds.
15	1870-1915	40	100	Algal unit: dark gray, cream-buff weathered carbonaceous dolomite; lower 25' pisolitic; upper 15' contains columnar stromatolites with heads 6-12" in diameter. Intraformational conglomerate between algal heads at top of unit. Bedding 105;55 N.
14	1790-1870	70	100	Siliceous dolomite: gray, gray weathered; massive; finely crystalline; abundant veinlets and irregular pods of quartz; some chert nodules near top.

SECTION 1, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
14	1540-1790	230	100	Siliceous dolomite: dark gray, buff to brown weathered; very finely crystalline; undulatory stromatolites; scattered glassy quartz sand grains; abundant digitate stromatolites as much as 2 mm in diameter. Bedding 120;55 N.
	1380-1540	140	100	Siliceous dolomite: gray to dark gray, buff weathered; fairly massive, medium crystalline, faint planar laminations in some exposures; a few irregular plates of quartz parallel to bedding; abundant digitate stromatolites, 1-3 mm in diameter.
	1334-1345	10	100	Siliceous dolomite: dark gray, buff brown and buff gray weathered; intraformational conglomerate; many quartz veinlets.
	1235-1345	95	100	Dolomite: gray, buff weathered; finely crystalline; massive; quartz lenses and layers parallel to bedding.
	1120-1235	110		Covered.
13	1095-1120	25	100	Algal unit: dark gray, buff weathered carbonaceous dolomite; small columnar stromatolites 3-6" in diameter; occur in zones separated by well laminated gray dolomite; some intraformational conglomerates; generally at top of zones.
12	965-1095	125	80	Dolomite: gray, cream to dark gray weathered; fine to medium crystalline; massive; milky quartz plates parallel to bedding.
11	950-965	15	100	Algal unit: columnar stromatolites 6-10" in diameter; differentially weathered gray and buff laminations 1-2 cm thick. Bedding 120;80 NE.

SECTION 1, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
10	855-950	90	100	Dolomite: gray, buff-gray weathered; well-laminated; medium-grained; scattered oval chert nodules and a few veinlets of milky quartz.
9	790-855	60	100	Algal horizon: 10' of intraformational conglomerate at the base; 2-4" hemispherical stromatolites compose most of the remainder; the upper 15' is pisolitic (concentrically laminated sub-spherical bodies, 0.2-4 cm in diameter); most of the rock is dark gray, carbonaceous dolomite.
8	660-790	125	100	Siliceous dolomite: gray, cream to gray weathered; fine to medium crystalline; a few quartz sand grains and irregular pods of quartz.
	555-660	105	100	Dolomite: light, gray, buff-gray weathered; finely crystalline; well-laminated in places; scattered quartz layers and lenses.
	500-555	55	100	Dolomite: light gray, dark weathered; medium crystalline, massive; a few quartz layers.
6	480-500	20	100	Dolomite: gray on fresh and weathered surfaces; fine to medium crystalline; scattered black chert nodules, elongate parallel to bedding. Bedding 130;75 N.
5	390-480	90	100	Dolomite: dark gray, buff-gray to dark gray weathered; undulatory stromatolites, but generally massive; many quartz veinlets parallel to bedding and fractures; chert nodules in the lower part. Bedding 115;V. Intraformational conglomerate at 465-480
	340-390	50	60	
	310-340	30	30	
3	110-210	100	20	Dolomite: dark gray; dove gray to tan-brown weathered; medium crystalline; undulatory stromatolites; milky quartz pods and veinlets. Bedding 130;V. Fractures at 15;V and 65;75 NW.

SECTION 1, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
2	110-210	100	20	Siliceous dolomite: cream gray, buff weathered; very well-laminated, finely crystalline; planar, closely spaced bedding surfaces; a few cross-beds.
	0 - 110	110		Covered.
Total Thickness		2580		

SECTION 2

Unit	Measured Distance	True Thickness	Percent Exposure	Description
19	2265-2315	40	100	Siliceous dolomite: cream gray to gray, buff-brown to tan weathered; very finely crystalline; very well-laminated; some chert nodules. Bedding 45;60 N.
18	2260-2265	20	100	Siliceous dolomite: gray, buff weathered; fine to medium crystalline; some undulatory stromatolites: intraformational conglomerate at base.
17	2220-2260	30	80	Algal unit: gray, buff weathered, slightly carbonaceous dolomite; 2' pisolitic layer overlain by columnar stromatolites, 5-8" in diameter.
16	2145-2220	60	80	Siliceous dolomite: gray, buff weathered; fine to medium crystalline; many irregular plates of quartz parallel to bedding and quartz veinlets along fractures; undulatory stromatolites in upper 25'.

SECTION 2, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
15	2130-2145	15	80	Algal unit: dark gray, buff weathered, medium crystalline carbonaceous dolomite; well-laminated to massive; many pisolitic layers; digitate stromatolites at 2005 and 2055.
	2055-2130	70	30	
	2005-2055	45	70	
14	1805-2005	185	20	Siliceous dolomite: gray, gray to buff weathered; finely crystalline, some undulatory and digitate stromatolites and some pisolites as much as 3 cm in diameter; quartz veinlets along fractures in upper 60'. Bedding 40;55 N.
	1705-1805	90	30	
	1540-1705	150		Covered.
	1330-1540	190	50	Siliceous dolomite: tan to light gray, buff weathered; very finely crystalline; massive; minor lenses and plates of milky quartz; a few pisolites and small digitate stromatolites towards top.
13	1305-1330	25	50	Algal unit: columnar stromatolites as much as 2" in diameter; buff to gray weathered carbonaceous dolomite. Bedding 50;60 NE.
12	1255-1305	45	50	Siliceous dolomite: gray to pink, dove gray to buff weathered; very finely crystalline; massive to finely bedded; some undulatory stromatolites; a few quartz lenses and chert nodules.
	1230-1255	25	100	
11	1190-1230	40	100	Algal unit: gray, buff weathered siliceous dolomite; lower 15' pisolitic; upper part contains many columnar stromatolites as much as 6-8" in diameter; some milky quartz along laminations.

SECTION 2, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
10	995-1190	180	40	Siliceous dolomite: light gray to cream, buff-gray weathered; fine to medium crystalline; well-laminated to massive, with plates and veinlets of quartz; some undulatory stromatolites at 1040-1060; a few chert nodules in upper part. Bedding 50;65 N.
	970-995	25	20	
9	935-970	30	20	Algal unit: columnar stromatolites, 4-7" in diameter and 2-4" high; dark gray, carbonaceous dolomite, fragmental at top.
8	895-935	30	20	Siliceous dolomite: pink to light gray, cream-buff weathered; very finely crystalline; massive; a few quartz pods; some quartz sand in upper 60'.
	800-985	85	80	
7	720-800	70	100	Siliceous dolomite: gray, cream-buff weathered; finely crystalline; thinly laminated; small amount of quartz in veinlets and pods.
	670-720	50		Covered.
6	645-670	25	50	Siliceous dolomite: light gray, buff weathered; very finely crystalline; finely laminated; a few black chert nodules; parting parallel to bedding. Bedding 40;70 N.
5	600-645	40	50	Siliceous dolomite: gray to pink, buff to dark gray weathered; finely crystalline; massive to poorly bedded; a few quartz pods, and quartz veinlets along fractures in upper 40'.
	500-600	90	20	
3	400-500	85	80	Siliceous dolomite: same as for 110-375.
	375-400	30		Covered.

SECTION 2, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
3	230-375	135	50	Siliceous dolomite: light gray, gray-brown weathered; fine to medium crystalline; massive to poorly laminated; partly brecciated, with patches of white quartz along fractures; irregular bedding; a few quartz pods and chert nodules; several thin layers of intraformational conglomerate. Bedding 30;60 NW.
	110-230	110	100	
2	0 -110	95	100	Siliceous dolomite: light cream-gray, tan to orange-buff weathered; very finely crystalline; massive fractures; parting parallel to planar bedding; very well-laminated; some cross-bedding; well developed joints at right angles to bedding. Bedding 35;70 NW.
Total Thickness		2115		

SECTION 3

Unit	Measured Distance	True Thickness	Percent Exposure	Description
19	915-1000	80	100	Siliceous dolomite: light gray, tan weathered; planar, closely spaced laminations; very finely crystalline; massive on fractured surface; a few chert nodules. Bedding 120;75 NE.
18	840-915	70	100	Dolomite: gray, buff weathered; fine to medium crystalline; 2' intraformational conglomerate at base; upper part exhibits undulatory stromatolites.

SECTION 3, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
17	810-840	30	100	Algal unit: dark carbonaceous dolomite; columnar stromatolites, 5-7" in diameter, weathered in relief along alternate laminations of gray and buff dolomite; fragmental material between the closely packed heads.
16	510-810	295	20	Dolomite: gray, light gray weathered; medium to finely crystalline; abundant plates of milky quartz parallel to bedding and along fractures; undulatory stromatolites in the upper 90'. Bedding 115;85 NE.
	440-510	70	80	Dolomite: gray, gray to buff weathered; massive; a few ovoid chert nodules and milky quartz pods; many quartz veinlets along fractures.
15	390-440	50	80	Algal unit: dark carbonaceous dolomite; 5' basal intraformational conglomerate; columnar stromatolites above this; the largest heads, 8-10" in diameter, are at the top; well-laminated dolomite separates successive colonies. Bedding 110;V.
14	305-360	55	80	Siliceous dolomite: dark gray, gray to buff weathered; finely crystalline; a few chert nodules and quartz pods; many quartz veinlets along fractures.
	195-305	105	100	Siliceous dolomite: dark gray; cream to dark gray weathered; slightly carbonaceous; undulatory and digitate stromatolites compose about 20 percent of the unit; pisolitic from 270-285. Bedding 115;80 NE.
	0 -195	190	100	Siliceous dolomite; light gray, cream-buff to dark gray weathered; massive, finely crystalline; a few milky quartz lenses parallel to bedding, where present; intraformational conglomerates at 85-90 and 120-135.
Total Thickness		945		

SECTION 4

Unit	Measured Distance	True Thickness	Percent Exposure	Description
17	2255-2295	35	75	Algal unit: gray, light gray weathered carbonaceous dolomite; finely crystalline; concentrically laminated pisolites 0.2-1 cm in diameter, and many digitate stromatolites in lower part; some 6-19" columnar stromatolites at top.
16	2040-2255	205	75	Dolomite: gray, light gray weathered, finely crystalline; massive; a few chert nodules and thin quartz-carbonate veins; quartz lenses abundant near top. Bedding 105;55 N.
15	1965-2040	70	85	Algal unit: concentric pisolites, 0.2-2 cm in diameter in a gray, buff weathered, fine-grained matrix; 8-10" columnar stromatolites in upper 15' of unit; a few digitate stromatolites in the lower part.
14	1945-1965	20	85	Siliceous dolomite: gray, cream-buff weathered, very fine-grained; massive; undulatory stromatolites near base; several zones of pisolites and digitate stromatolites. Quartz and carbonate veins abundant. Scattered quartz lenses in central part and chert nodules near top.
	1856-1945	75	25	
	1830-1865	35	45	Dolomite: gray, buff weathered; finely crystalline; undulatory stromatolites; pisolites in some exposures. Bedding 100;70 N.
	1775-1830	50		Covered.
	1720-1775	55	75	Dolomite: dark gray, dark weathered; undulatory laminations near top; quartz veins in massive lower part; pisolites and 1" digitate stromatolites at 1740; some intraformational conglomerates.
	1675-1720	40		Covered.

SECTION 4, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
14	1590-1675	70	25	Dolomite: gray, dove-gray to buff weathered; massive at base, some undulatory stromatolites at top; many fragmental zones, and some pisolites at 1555; a few chalcedony pods and quartz lenses.
	1535-1590	50	100	
	1325-1535	185		Covered.
	1215-1325	95	70	Dolomite: gray, dove-gray to dull buff weathered, finely crystalline, massive; quartz stringers and pods accentuate the poor bedding. Bedding 100;70 N. Pisolites in upper 10' of exposure, 1-3 cm in diameter, concentrically banded.
13	1205-1215	10	70	Algal unit: dark carbonaceous dolomite; columnar stromatolites, 2-6" in diameter.
12	1095-1205	90	25	Dolomite: cream-gray, tan weathered; medium to finely crystalline; massive; a few quartz lenses.
	1075-1090	15	85	
11	1060-1075	15	85	Algal unit: columnar stromatolites 6-8" in diameter, overlain by intraformational conglomerate. Dolomite is tan to flesh-colored, buff weathered.
10	1035-1060	20	85	Dolomite: gray, buff weathered, massive; a few chert nodules and quartz-carbonate stringers. Bedding 85;65 N.
	975-1035	55	55	
	925-975	45	85	
9	900-925	25	85	Algal unit: cream gray, siliceous dolomite, light brown weathered, massive, columnar stromatolites 2-4" wide in lower 15'. Subspherical pisolites in upper part.
8	750-900	135	75	Siliceous dolomite: dark gray, buff to dark weathered; massive; finely crystalline. A few quartz grains in some exposures. Some chalcedony pods.

SECTION 4, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
8	640-750	95	40	Siliceous dolomite: gray, dove-gray weathered, finely crystalline; a few irregular pods of chalcedony, and veinlets of quartz; poor bedding. Bedding 105;65 NE.
	525-640	100	20	Siliceous dolomite: gray brown, buff brown weathered; finely crystalline, massive; many pink and white quartz-carbonate veinlets and coarsely crystalline patches of quartz.
6	485-525	30	40	Dolomite: gray, gray-buff weathered; medium to finely crystalline. A few ovoid black chert nodules, flattened parallel to bedding.
5	350-485	120	40	Siliceous dolomite: gray, light buff to gray weathered; dark brown beneath moss; massive, finely crystalline; several flat pebble conglomerates; a few chert nodules and numerous small carbonate stringers. Increasingly carbonaceous towards top, where undulatory stromatolites are prominent. Exposures interrupted by swamps. Bedding 100;60N.
4	335-350	15	40	Siliceous dolomite and chert: dark gray to black chert fragments in a light gray siliceous matrix.
3	280-335	50		Covered.
	205-280	70	20	Dolomite: gray, buff weathered, carbonaceous; finely crystalline; lowermost 20' massive with a few quartz-filled fractures (35;60 NW), undulatory stromatolites and pisolites in upper part. Bedding 95;55 N.
2	100-180	70	30	Siliceous dolomite: cream-buff; orange-buff weathered; very finely crystalline; very well-laminated; planar beds less than 5 mm thick are clearly visible on ribbed weathered surfaces, but not visible on smooth-fracturing fresh surfaces; small-scale cross-bedding. Bedding 95;60 N.

SECTION 4, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
2	0 -100	90		Covered.
Total Thickness		2055		

SECTION 5

Unit	Measured Distance	True Thickness	Percent Exposure	Description
18	2885-2945	50	100	Dolomite: gray, gray weathered; undulatory stromatolites; finely to medium crystalline; smooth fractures.
17	2875-2885	10	80	Algal unit: carbonaceous dolomite; columnar stromatolites, 6-10" in diameter; alternate buff and gray laminations.
16	2835-2875 2740-2835 2715-2740	40 90 30	60 20 80	Dolomite: gray, buff-gray weathered; medium to finely crystalline; many chert nodules and carbonate veins; lenses of quartz near the top.
15	2715-2735	15	60	Algal unit: dark carbonaceous dolomite; columnar stromatolites 8-12" in diameter; alternate buff and light gray laminations.
14	2650-2715	50	60	Siliceous dolomite: light gray, gray weathered; massive; scattered chert nodules and a few irregular pods of milky quartz; quartz stringers.
	2580-2650	55		Covered.

SECTION 5, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
14	2395-2580	175	20	Dolomite: gray, buff-gray weathered; finely to medium crystalline; undulatory stromatolites; algal pisolites at 2250-2255 and 2415-2430. Pisolites are sub-spherical, concentrically laminated, 0.2-3 cm in diameter. Bedding 90;70 N.
	2335-2395	60	40	
	2035-2335	290		Covered.
	1885-2035	145		Dolomite: same as unit 12.
13	1875-1885	10	80	Algal unit: buff to gray, carbonaceous dolomite; columnar stromatolites, 4-6" in diameter; poorly laminated.
12	1660-1875	215	80	Dolomite: dark gray, gray to tan weathered; medium crystalline; massive; scattered quartz stringers and lenses parallel to bedding. Bedding 85;75 N.
11	1650-1660	10	100	Algal unit: carbonaceous dolomite; columnar stromatolites 8-10" in diameter; alternate gray and buff laminations, differentially weathered.
10	1400-1650	245	100	Dolomite: light gray to buff-gray, cream-buff weathered; massive; rough weathered surface; many gray to white chert nodules; scattered irregular milky quartz pods and carbonate veinlets parallel to bedding and along fractures at 40;70 NW and 155;60 NE.
9	1385-1400	15	100	Algal unit: gray to buff carbonaceous dolomite; columnar stromatolites 4-8" in diameter, ribbed weathered surface outlines the laminations.
	1375-1385	10	100	Flat pebble conglomerate: dark gray, dark weathered fragments in an arenaceous dolomite matrix.

SECTION 5, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
8	1230-1375	145	80	Dolomite: gray to light gray, buff weathered; fine to medium crystalline; massive; scattered quartz grains; many irregular lenses of white quartz. Bedding 80;V.
	1205-1230	25	80	Siliceous dolomite breccia: light gray, medium grained to finely crystalline fragments cemented by milky quartz; massive; rough weathered surface; numerous irregular quartz pods and veinlets. Most prominent fractures at 150;V.
	1095-1205	110	90	Dolomite: light gray, dove gray weathered; medium crystalline; scattered quartz sand grains; rough weathered surface; obscure bedding; a few irregular pods of quartz; two prominent sets of intersecting carbonate veinlets. Bedding 85;V.
7	1030-1095	60	90	Dolomite: dark gray, gray weathered; medium to finely crystalline; well-laminated; a few irregular masses of milky quartz. Bedding 85;V.
6	1010-1030	20	90	Dolomite: dark gray, buff weathered; medium to finely crystalline; black chert; ovoid nodules elongate parallel to bedding.
5	970-1010	40	30	Dolomite: gray to dark gray; buff to gray weathered; medium to finely crystalline; undulatory stromatolites. Bedding 50;V.
	905-970	60	30	Dolomite: gray to dark gray; tan to gray weathered; medium to finely crystalline; massive; scattered gray to white chert nodules, ovoid in section, elongate parallel to bedding; numerous quartz-carbonate veinlets less than 2 cm thick follow fracture sets at 5;60 W and 120;75 NE.

SECTION 5, continued

Unit	Measured Distance	True Thickness	Percent Exposure	Description
4	890-905	15	60	Siliceous dolomite, and chert: dull black, rusty to dark weathered; much fractured and brecciated. Grades upward to green-gray, rusty gray weathered siliceous dolomite which exhibits faint bedding.
3	715-890	175	60	Siliceous dolomite: dark gray, buff to gray weathered; finely crystalline; bedding obscure; some undulatory stromatolites. Bedding 55;80 NW.
2	635-715		10	Dolomite: gray, cream-buff weathered; finely crystalline; well-laminated planar beds increasingly thinner and more regularly spaced towards the top. Local small-scale cross-bedding. Bedding 65;86 NW.
	565-635		30	
	465-565	235	100	
	240-465	215	0	Marion Lake.
	65-240	165	100	Siliceous dolomite: light gray, tan-buff weathered; very finely crystalline; very well-laminated, with ribbed weathered surface controlled by bedding; beds paper-thin to 1 mm thick, planar and remarkably persistent; local small-scale cross-bedding. Bedding 60;V.
1	0 - 65	60	100	Siliceous dolomite: cream-gray, orange to tan weathered, very finely crystalline; planar-bedded; fair to well-laminated; grid pattern of surface grooves developed by fracture-controlled weathering. Bedding 70;80 SE. Fracture sets at 10;45 W and 100;85 SW.
Total Thickness		2860		

BIBLIOGRAPHY

Andrichuk, J. M., 1955, Mississipian Madison group stratigraphy and sedimentation in Wyoming and southern Montana: Am. Assoc. Petroleum Geologists Bull., v. 39, p. 2170-2210

Bagnold, R. A., 1941, The physics of blown sand and desert dunes: 1st ed., London, Methuen and Co. Ltd., 265 p.

Barrell, Joseph, 1906, Relative geological importance of continental, littoral, and marine sedimentation, Part III: Jour. Geol., v. 14, p. 524-568

Belyea, H. R., and Scott, A. W., 1935, Conditions of sedimentation of the Halifax formation as observed in Point Pleasant Park: Nova Scotian Inst. of Sci. Proc., v. 18, p. 225-239

Biggs, D. L., 1957, Petrology and origin of Illinois nodular cherts: Ill. Geol. Survey Circ. 245, 25 p.

Bird, J. B., 1953, The glaciation of Central Keewatin, Northwest Territories, Canada: Am. Jour. Sci., v.251, p. 215-230

Black, Maurice, 1933, The algal sediments of Andros Island, Bahamas: Roy. Soc. London Philos. Trans., ser. B, v. 222, p. 165-192

Bradley, W. H., 1929, Algae reefs and oolites of the Green River formation: U. S. Geol. Survey Prof. Paper 154, p. 203-223

Bramlette, M. N., 1946, The Monterey formation of California and the origin of its siliceous rocks: U. S. Geol. Survey Prof. Paper 212, 57 p.

Bretz, H. J., and Horberg, Leland, 1949, Caliche in southeastern New Mexico: Jour. Geol., v. 57, p. 491-511

Bucher, W. H., 1919, On ripples and related sedimentary surface forms and their paleogeographic interpretation: Am. Jour. Sci., 4th ser., v. 47, pt. 1, p. 149-210

Chave, K. E., 1954, Aspects of biochemistry of magnesium: Jour. Geol., v. 62, pt. 1, p. 266-283

Chayes, Felix, 1949, A simple point counter for thin-section analysis: Am. Mineralogist, v. 34, p. 1-11

Chilingar, G. V., 1956, Relationship between Ca-Mg ratio and geologic age: Am. Assoc. Petroleum Geologists Bull., v. 40, no. 9, p. 2256-2266

Clarke, J. M., 1900, The water biscuit of Squaw Island, Canandaigua Lake, N. Y.: N. Y. State Mus. Bull., no. 39, v. 8, p. 195-198

Cloos, Ernst, 1950, The geology of the South Mountain anticlinorium: The Johns Hopkins University Studies in Geology, no. 16, Guidebook I, 28 p.

Cloud, P. E., Jr., 1942, Notes on stromatolites: Am. Jour. Sci., v. 240, p. 363-379

_____, 1945, The stromatolite gymnoscolen not a salinity index: Am. Jour. Sci., v. 243, no. 2, p. 108

Collins, W. H., 1925, North Shore of Lake Huron: Canada Geol. Survey Mem., v. 143, 160 p.

Correns, C. W., 1950, The geochemistry of diagenesis: Geochim. et Cosmochim. Acta, v. 1, p. 49-54

Dake, C. L., 1921, The problem of the St. Peter sandstone: Mo. School of Mines Bull., v. 6, no. 1, 228 p.

Davis, E. F., 1918, The radiolarian cherts of the Franciscan group: California Univ. Pub. Geol. Sci. Bull., v. 11, p. 235-432

De Geer, Gerard, 1912, A geochronology of the last 12,000 years: Congr. géol. Intern., Sess. 11 C. R., 1910, p. 241-258

De Sitter, L. V., 1956, Structural geology, New York, McGraw-Hill Co., 552 p.

Duffell, Stanley, 1959, Mount Wright map-area, Quebec: Canada Geol. Survey Paper 59-6

Eardley, A. J., 1938, Sediments of Great Salt Lake, Utah: Am. Assoc. Petroleum Geologists, v. 22, p. 1305-1411

Edie, R. W., 1956, Origin and characteristics of evaporitic dolomite: Alberta Soc. Petroleum Geologists Jour., v. 4, no. 1, p. 16-23

Emig, W. H., 1917, Travertine deposits of Oklahoma: Oklahoma Geol. Survey Bull. 29, 76 p.

Emmons, R. C., 1943, The universal stage: Geol. Soc. America Mem. 8, 205 p.

Fahrig, W. F., 1951, Griffis Lake, Quebec: Canada Geol. Survey Paper 51-23

_____, 1953, Griffis Lake area, New Quebec: Unpub. Ph.D. thesis, University of Chicago

_____, 1955, Lac Herodier, New Quebec: Canada Geol. Survey Paper 55-1

_____, 1957, Geology of certain Proterozoic rocks in Quebec and Labrador: Roy. Soc. Canada Special Pub. no. 2, p. 112-123

Fairbridge, R. W., 1946, Submarine slumping and location of oil bodies: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 84-92

Fairbridge, R. W., 1957, The dolomite question: Soc. Econ. Paleontologists and Mineralogists, Spec. Pub., no. 5, p. 125-178

Field, R. M., 1916, A preliminary paper on the origin and classification of intraformational conglomerates: Ottawa Naturalist, v. 30, nos. 2-6, p. 29-36, 47-52, 58-66

Fleming, R. H., and Revelle, Roger, 1939, Physical processes in the ocean; Recent marine sediments: Soc. Econ. Paleontologists and Mineralogists, Spec. Pub., no. 4, p. 95-102

Folk, R. L., 1959, Practical petrographic classification of limestones: Am. Assoc. Petroleum Geologists, v. 43, no. 1, p. 1-38

Frarey, M. J., 1952, Willbob Lake, Quebec and Labrador: Canada Geol. Survey Paper 52-16

_____, 1954, Geology of the Willbob Lake area, Northern Quebec and Western Labrador: Unpub. Ph.D. thesis, University of Michigan

Friedman, G. M., 1958, Determination of sieve-size distribution from thin-section data for sedimentary petrological studies: Jour. Geol., v. 66, p. 394-416

Ginsburg, R. N., 1958, The influence of marine bottom communities on the depositional environment of sediments: Jour. Geol., v. 66, p. 310-318

Glagolev, A. A., 1933, On geometrical methods of quantitative mineralogical analysis of rocks: Atomic Energy Research Establishment Library, Translation 567, Harwell, Berks, 1955

Gordon, M. and Tracey, J. I., 1952, Origin of the Arkansas bauxite deposits, in: Problems of clay and laterite genesis: Am. Inst. Mining Engineers Symposium, New York, p. 12-34

Greenman, N. N., 1951, The mechanical analysis of sediments from thin-section data: Jour. Geol., v. 59, p. 447-462

Gross, G. A., 1951, Comparative study of three slate formations of the Ferriman series in the Labrador Trough: Unpub. M.Sc. thesis, Queen's University, Canada

Grout, F. F., and Schwartz, G. M., 1933, The geology of the Rove formation and associated intrusives in northeastern Minnesota: Minn. Geol. Survey Bull. 24, 103 p.

Haff, J. C., 1938, Preparation of petrofabric diagrams: Am. Mineralogist, v. 23, p. 543-574

Harrison, J. M., 1952, The Quebec-Labrador iron belt, Quebec and Newfoundland (Preliminary Report): Canada Geol. Survey Paper 52-20

Hobbs, C. R. B., Jr., 1957, Petrography and origin of dolomite-bearing carbonate rocks of Ordovician age in Virginia: Va. Polytech. Inst., v. 50, no. 5, 128 p.

Illing, L. V., 1954, Bahaman calcareous sands: Am. Assoc. Petroleum Geologists Bull., v. 38, no. 1, p. 1-95

James, H. L., 1951, Iron formation and associated rocks in the Iron River district, Iron County, Michigan: Geol. Soc. America Bull., v. 62, p. 251-266

Johnson, J. H., 1946, Lime-secreting algae from the Pennsylvanian and Permian of Kansas: Geol. Soc. America Bull., v. 57, p. 1087-1120

Jones, J. C., 1925, The geologic history of Lake Lahontan: Carnegie Inst. Washington Pub. 352, p. 1-50

Kalkowsky, Ernst, 1908, Oolith und Stromatolith in Norddeutschen Buntsandstein: Deutsch geol. Gesell. Zeitschr., v. 60, p. 68-125

Knight, S. H., 1930, Festoon cross-lamination (abs.): Geol. Soc. America Bull., v. 41, p. 86

Krumbein, W. C., 1933. Textural and lithologic variations in glacial till: Jour. Geol., v. 41, p. 382-408

_____, 1935, Thin-section mechanical analysis of indurated sediments: Jour. Geol., v. 43, p. 482-496

_____, 1940, Flood gravel of San Gabriel Canyon, California: Geol. Soc. America Bull., v. 51, p. 639-676

_____, 1942, Flood deposits of Arroyo seco, Los Angeles County, California: Geol. Soc. America Bull., v. 53, p. 1355-1402

_____, and Garrels, R. M., 1952, Origin and classification of chemical sediments in terms of pH and oxidation-reduction potentials: Jour. Geol., v. 60, p. 1-33

_____, and Pettijohn, F. J., 1938, Manual of sedimentary petrography: 1st ed., New York, Appleton-Century-Crofts Inc., 549 p.

_____, and Sloss, L. L., 1953, Stratigraphy and sedimentation: San Francisco, W. H. Freeman and Co. Ltd., 497 p.

Krynine, P. D., 1940, Petrology and genesis of the Third Bradford sand: Pa. State Coll. Bull. 29, 134 p.

_____, 1943, Diastrophism and the evolution of sedimentary rocks: A lecture sponsored by Am. Assoc. Petroleum Geologists, State College, Pa., 21 p.

Kuenen, P. H., and Migliorini, C. I., 1950, Turbidity currents as a cause of graded bedding: Jour. Geol., v. 58, p. 91-127

Kupferburger, W., 1935, The travertine deposits near Port St. Johns: South Africa Dept. Mines Geol. Series, Bull. 4, 15 p.

Long, M. B., 1955, Origin of the Conococheague limestone: Unpub. Ph.D. thesis, The Johns Hopkins University

Low, A. P., 1895, Report on exploration in the Labrador peninsula along the Eastmain, Koksoak, Hamilton, Manicuagan and portions of other rivers: Canada Geol. Survey, Ann. Rept. L, new series, v. VIII, 387 p.

McKee, E. D., 1953, Report on studies of stratification in modern sediments and in laboratory experiments: Office of Naval Research, Proj. Nonr 64 (00), NR 081 123, 61 p.

Mann, V. I., 1951, Spot test for phosphorous in rocks: Jour. Sed. Pet., v. 20, no. 2, p. 116-117

Mansfield, G. R., 1927, Geography, geology and mineral resources of part of southeastern Idaho: U. S. Geol. Survey Prof. Paper 152, 453 p.

Moore, D. G., and Scruton, P. C., 1957, Minor internal structures of some recent unconsolidated sediments: Am. Assoc. Petroleum Geologists Bull., v. 41, p. 2723-2751

Murray, George, 1895, Calcareous pebbles formed by algae: Phycological Memoirs, p. 74-77

Nockolds, S. R., 1954, Average chemical composition of some igneous rocks: Geol. Soc. America Bull., v. 65, p. 1007-1032

Ohle, E. L., and Brown, J. S., 1954, Geologic problems in the southeast Missouri Lead District: Geol. Soc. America Bull., v. 65, p. 201-222

_____, 1954, Geologic problems in the southeast Missouri Lead District, Supplement: Geol. Soc. America Bull., v. 65, p. 935-936

Pelletier, B. R., 1957, Pocono paleocurrents: Unpub. Ph.D. thesis, The Johns Hopkins University

_____, 1958, Pocono paleocurrents in Pennsylvania and Maryland: Geol. Soc. America Bull., v. 69, p. 1033-1064

Pettijohn, F. J., 1936, An early Pre-Cambrian varved slate in northwestern Ontario: Geol. Soc. America Bull., v. 47, p. 621-628

_____, 1957, Sedimentary rocks: 2nd ed., New York, Harper and Brothers, 718 p.

Pia, Julius, 1933, Die Rezenten Kalksteine: Zeitschr. für Kristalographie, Mineralogie, und Petrographie Abt. B, p. 1-420

Plumley, W. J., 1948, Black Hills terrace gravels; a study in sediment transport: Jour. Geol., v. 56, p. 526-577

Potter, E. P., and Glass, H. D., 1958, Petrology and sedimentation of the Pennsylvanian sediments in southern Illinois; a vertical profile: Ill. State Geol. Survey Rept. Inv. 204, 60 p.

Price, W. A., Elias, M. K., and Frye, J. C., 1946, Algae reefs in caprock of Ogallala formation on Llano Estacado Plateau, New Mexico and Texas: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 1742-1746

Rezak, Richard, 1957, Stromatolites of the Belt Series in Glacier National Park and vicinity, Montana: U. S. Geol. Survey Prof. Paper 294-D, p. 127-151

Rich, J. L., 1951, Probable fondo origin of Marcellus-Ohio-New Albany-Chattanooga bituminous shales: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 2017-2040

Roddy, H. J., 1915, Concretions in streams formed by the agency of blue-green algae and related plants: Am. Philos. Soc., v. 54, p. 246-258

Roscoe, S. M., 1957, Cambrian Lake (east half), New Quebec: Canada Geol. Survey Paper 57-6

Russell, J. C., 1885, Geological history of Lake Lahontan: U. S. Geol. Survey Mon., v. 11, p. 1-288

Sander, Bruno, 1926, Zur petrographisch-tektonischen Analyse III: Jahrbuch d. B. A., p. 323-406

Sando, W. J., 1957, Beekmantown group (Lower Ordovician) of Maryland: Geol. Soc. America Mem. 68, 161 p.

Sauramo, Matti, 1923, Studies on the Quarternary varve sediments in southern Finland: Comm. Geol. Finlande, Bull. no. 60, 164 p.

Sauvé, Pierre, 1957, The geology of the east half of the Gerido Lake area, New Quebec, Canada: Unpub. Ph.D. thesis, The Johns Hopkins University

Schmidt, R. G., 1957, Joint patterns in relation to regional and local structure in the Central foothills belt of the Rocky Mountains of Alberta: Unpub. Ph.D. thesis, University of Cincinnati, 185 p.

Schneiderhöhn, H., 1921, Beitrage zur Kenntis der Erzlagerstatten und der geologischen Verhaltnisse des Otavi berglandes, Deutsch-Sudwestafrika: Abh. senchenb. naturf. Ges., v. 37, no. 3, p. 264

Shrock, R. R., 1948, Sequence in layered rocks: New York, McGraw Hill, 507 p.

Siever, Raymond, 1949, Trivoli sandstone of Williamson County, Illinois: Jour. Geol., v. 57, p. 614-617

Simonen, Ahti, and Kouvo, Olavi, 1951, Archean varved schists north of Tampere in Finland: Bull. Comm. Geol. Finn., no. 154, p. 93-113

Sproule, J. C., 1939, The Pleistocene geology of the Cree Lake region Saskatchewan: Roy. Soc. Canada Trans., ser. 3, sec. 4, p. 101

Strakhov, N. M., 1953, Diagenesis of sediments and its significance for sedimentary ore formation: S.S.S.R. Akad. Nauk. Izv., Ser. Geol., no. 5, p. 12-49

Swineford, Ada, Byron, L. A., and Frye, J. C., 1958, Petrology of the Pliocene pisolitic limestone in the Great Plains: Kansas State Geol. Survey Bull. 130, pt. 2, p. 97-116

Taliaferro, N. L., 1933, The relation of volcanism to diatomaceous and associated siliceous sediments: California University Pub. Geol. Sci. Bull., v. 23, p. 1-55

Thiel, G. A., 1935, Sedimentary and petrographic analysis of the St. Peter sandstones: Geol. Soc. America Bull., v. 46, p. 556-614

Thompson, W. O., 1937, Original structures of beaches, bars, and dunes: Geol. Soc. America Bull., v. 48, p. 723-752

Trask, P. D., 1932, Origin and environment of source sediments of petroleum: Houston, Texas, Gulf Publishing Co., p. 71-72

Turner, F. J., 1948, Mineralogical and structural evolution of the metamorphic rocks: Geol. Soc. America Mem. 30, 342 p.

Twenhofel, W. H., 1919, Pre-Cambrian and carboniferous algal deposits: Am. Jour. Sci., v. 48, p. 339-352

_____, 1932, Treatise on sedimentation: 2nd ed., Baltimore, The Williams and Wilkins Company, 926 p.

_____, 1939, Environments of origin of black shales: Am. Assoc. Petroleum Geologists Bull., v. 23, p. 1178-1198

_____, 1945, The rounding of sand grains: Jour. Sed. Pet., v. 15, no. 2, p. 59-71

Udden, J. A., 1914, Mechanical composition of clastic sediments: Geol. Soc. America Bull., v. 25, p. 655-744

Van Hise, C. R., and Leith, C. K., 1911, The geology of the Lake Superior region: U. S. Geol. Survey Monograph 52, 641 p.

Van Straaten, L. M. J. U., 1951, Texture and genesis of Dutch Wadden Sea sediments: Int. Cong. Sedimentology, 3rd Proc., p. 225-244

Van Tuyl, F. M., 1914, The origin of dolomite: Iowa Geol. Survey Ann. Rept., v. 25, p. 251-422

Walcott, C. D., 1894, Paleozoic intraformational conglomerates: Geol. Soc. America Bull., v. 5, p. 191-198

Wentworth, C. K., 1932, The mechanical composition of sediments in graphic form: Iowa University Studies in Natural History, v. 14, no. 3, 127 p.

Wilson, M. E., 1941, Noranda district, Quebec: Canada Geol. Survey Mem. 229, 162 p.

Winchell, A. N., 1951, Elements of optical mineralogy, part 2, description of minerals: New York, John Wiley and Sons, 4th ed., 551 p.

Yoder, H. S., and Eugster, H. P., 1955, Synthetic and natural muscovites: Geochim. et Cosmochim. Acta, v. 8, no. 5/6, p. 225-280

Young, R. B., 1943, The domical-columnar structure and other minor deformations in the Dolomite Series: South Africa Geol. Soc. Trans., v. 46, p. 91-106

VITA

John Allan Donaldson was born October 15, 1933, in Chatham, Ontario, where he attended McKeough Public School and Chatham Collegiate Institute. He entered Queen's University in September, 1952 and graduated with the degree of B.Sc. in 1956. His thesis study at Queen's University was: "Petrogenesis of a Lithium-bearing Pegmatite Dyke." In September, 1956 he began post graduate study at The Johns Hopkins University, Baltimore, Maryland. An Instructorship was awarded for the academic year 1956-57, and Student Assistantships were awarded for the years 1957-58 and 1958-59. A Gilman Scholarship was held during the three years of residence at Baltimore.

Donaldson has had the following field experience: 1953, Geological Survey of Canada, junior assistant engaged in four-mile reconnaissance mapping, Ungava; 1954, Cartier Mining Corporation, senior assistant, mapping iron ore deposits, Quebec; 1955, Imperial Oil Company of Canada, geological trainee engaged in well-site work and stratigraphic studies, southwestern Ontario; 1956, British Canadian Lithium Corporation, resident geologist, mapping and supervising exploratory diamond drilling of pegmatites; 1957, 1958, Geological Survey of Canada, party chief, engaged in one-mile mapping, Labrador and Quebec; 1959, Geological Survey of Canada, party chief, engaged in four-mile reconnaissance mapping, northwestern Ontario.

Donaldson is a member of Sigma Xi, Geological Association of Canada, Association of Professional Engineers of Ontario, and Canadian Institute of Mining and Metallurgy.

In 1957 Donaldson married Elizabeth Maxwell Grimmer of St. Andrews, New Brunswick. He joined the Geological Survey of Canada in June, 1959, and lives now in Ottawa, Ontario.

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